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DIFFERENTIATION IN XENOLITHIC LAMPROPHYRE DIKES AT MARQUETTE, MICHIGAN

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ABSTRACT

A fine network of aplites, pegmatites, and veins, together with partially assimilated granitic xenoliths, reveals the progress of differentiation within a set of lamprophyric dikes. First, intratelluric augite and biotite were replaced by pargasite, with simultaneous precipitation of more pargasite and of intermediate plagioclase and saussurite. These were followed by albite as aplite, by microcline and muscovite as micropegmatite, and by epidote and calcite as veins. A quantitative analysis shows that the dike rock corresponds to Camptonose III.I.5.3.4, and suggests that the average amphibole is low in lime and perhaps high in potash.

INTRODUCTION

Three lamprophyre dikes here described are located in the city of Marquette, Michigan, along North Sixth Street at the extension of Hewett Avenue. They are in the central part of the NE. $\frac{1}{4}$ of NW. $\frac{1}{4}$ of NW. $\frac{1}{4}$ of Sec. 23, T. 48 N., R. 25 W. They are part of a low *roche moutonnée*, consisting principally of banded greenstone and projecting through a cover of old beach sands which are now 140 feet above the level of Lake Superior. A plan of the outcrop is shown in Figure 1. Similar xenolithic lamprophyre dikes are found at intervals eastward for $1\frac{1}{4}$ miles to Light House Point; another dike is located on Middle Island Point, 3 miles to the north. The largest of such xenolithic dikes in Marquette County known to the writers is 50 feet wide; it lies about 9 miles west-northwest of the exposure under discussion. It there intersects a conglomeratic slate, which is

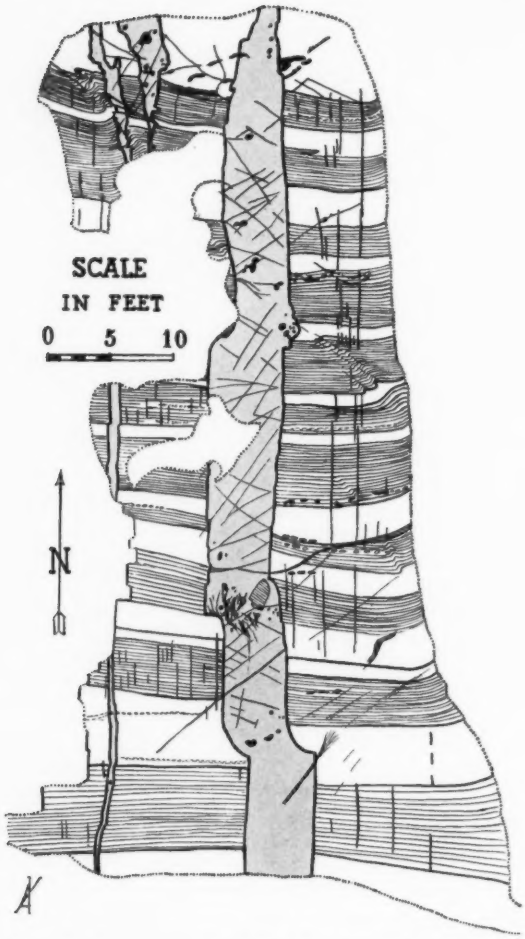

FIG. 1.—Map of rock outcrop at North Sixth Street and Hewett Avenue, Marquette, Michigan. Greenstone cut by spessartite dikes (shaded) with enclosed granitic xenoliths (black). *E–W* bedding planes and laminations in the greenstone, as well as *N–S* and oblique joint planes, are represented by solid lines; *E–W* ferruginous layers by parallel broken lines; pillow lava layers by strings of small ovals; and vein quartz in the greenstone by solid black areas.

presumably basal Upper Huronian and which is part of the north limb of a trough of Huronian rocks known as the Dead River basin.[1]

GENERAL GEOLOGIC RELATIONS

The rock intruded by the lamprophyre dikes at North Sixth Street is the Mona greenstone schist, usually regarded as Keewatin. It is in general so fine grained that only a small percentage of its constituent minerals can be identified with a hand lens; consequently, structure rather than composition furnishes its distinguishing features in the field. Actually, the Mona greenstone is characterized by two types of structure—massive and laminated—which alternate at irregular intervals. The massive beds are dark green in color, due to microscopic chlorite and actinolite; the laminated beds are intercalations of dark- and light-green stripes, the lighter ones being due to an admixture of saussurite. Cherty or heavily iron-stained layers furnish an occasional modification; one 4-cm. bed of fairly uniform ellipsoids is probably a variation of pillow lava.[2] The strike of the beds is nearly east-west. The dip is to the north and rolls between 70° and 80°. Vertical north-south trending joints, spaced from a millimeter to several meters apart, intersect the laminated portions of the greenstone and also the thinner of the massive beds; they enter certain of the thicker flows but do not completely cross them. Into this system of joints have been intruded the dikes under discussion.

Most of the lamprophyre in the exposure is of the variety known as "spessartite." The dikes have two characteristics by which they may be correlated with contemporaries in the immediate vicinity and by which they may be distinguished from others of similar lithology but of different age. The first and more important of the characteristics is the presence of rounded granitic xenoliths up to 20 cm. in diameter; the other is the relationship between the dikes and discontinuous joints of north-south trend. The joints are interrupted in the massive greenstone layers; the dikes likewise are in-

[1] C. R. Van Hise and C. K. Leith, "Geology of the Lake Superior Region," *U.S. Geol. Surv. Mono. 52* (1911), p. 287, with map by A. E. Seaman.

[2] In a pioneer study Williams interpreted these greenstones as representing metamorphosed lava flows with interbedded volcanic ash (G. H. Williams, "The Greenstone Schist Area of the Menominee and Marquette Regions of Michigan," *U.S. Geol. Surv. Bull. 62* [1892]).

terrupted, and usually the separated portions are offset to the east or west. As a rule the dikes pinch out in the massive beds; but occasionally the lamprophyres appear at the surface to have been cut across cleanly by a thick greenstone layer which resembles a sill and thus seems younger than the north-south dikes. Where the exposure can be seen in three dimensions rather than in two, however, the "sill" is shown to be just a concordant greenstone bed with "greenstick" fracture through which the dike weaves in such a way that it does not appear at all in some of the surface sections.

The source of the xenoliths is believed to have been granite, a phase of which is apparently represented in the locality in sills of sheared rhyolitic felsophyre intruding the greenstone. To the north these sills increase in number, in size, and in coarseness of grain and become true granites within a distance of 2 miles. Their age is considered to be Laurentian by Van Hise and Leith.[3] Pebbles of similar nature, both granite and felsophyre, are to be found in a conglomeratic slate on the north limb of the Dead River basin, particularly in the S. $\frac{1}{4}$, Sec. 6, T. 48 N., R. 26 W. It has not been ascertained as yet whether this conglomeratic slate is basal Upper or basal Middle Huronian, or even, perhaps, Keewatin. In no case, however, can the granite which furnished the pebbles be as young as the post-Huronian Killarnean.

A precise determination of the age of the spessartites in the city of Marquette is not possible because of the absence of suitable criteria; however, a tentative correlation with the Dead River xenolithic lamprophyre and with the Clarksburg pyroclastics may be made. The lamprophyre north of Dead River trends northeast across the southward dipping conglomeratic slate mentioned above, but it is not known to cut the younger black Michigamme slate (Upper Huronian). It is thus dated as pre-Michigamme. In the Marquette trough, a few miles to the south, where the stratigraphy has been worked out in some detail, the only Huronian pyroclastic formation is the Clarksburg (pre-Michigamme Upper Huronian).[4] Like the lamprophyre, the Clarksburg consists of predominant amphibole with rare palimpsest augite and includes blocks of granite among its ejectamenta.

[3] *Op. cit.*, p. 255. [4] *Ibid.*, p. 268.

Incidentally, the diaschistic implication of the term "lamprophyre" must not be overlooked when the age possibilities of near-by granites are being considered, but no such relationship is probable between the spessartites and the granite of the xenoliths.

In the Sixth Street outcrop there occur, besides the greenstone and the dikes, two distinct sets of veins. The older veins consist of pinkish quartz in isolated lenses, trend north of east, and are intersected by the spessartite. The other set contains epidote with comb structure interfingered with later calcite, strikes N. 60° E., and cuts the spessartite. The latter veins are comparatively wide (up to 3 cm.) in the dike but branch out into scores of veinlets upon passing into the massive greenstone.

PETROGRAPHY OF THE SPESSARTITE

In the area mapped three facies of lamprophyric rocks may be distinguished petrographically. *Spessartite* forms the greatest part of the dikes exposed, *hornblendite* occurs as thin selvages on the margins, and *mica lamprophyre*, or *malchite*, predominates at the south end of the westernmost dike.

Slightly over half of the spessartite (Fig. 2) consists of green amphiboles, occurring as

1. Sharp idiomorphs with (110) and (010) faces, but without good terminal planes, many having simple contact twinning parallel to (100). The X vibration direction is straw yellow to nearly colorless, Y is light greenish brown, and Z is green. α is 1.653, β is 1.663, and γ is 1.675, all ± 0.003; $2V$ (+) is $85° \pm 2°$; $Z\wedge c$ is $26° \pm 2°$. These are the properties of a "pargasite." In thickness the individual crystals range from 0.02 mm. at the margins of the dikes to 0.5 mm. in the interior.

The largest crystals show distinct zoning. On the interior is the pargasite just described. The next layer is also an optically positive pargasite, with a birefringence, however, of only 0.017. It is almost colorless on the inside, the color intensity increasing outward to light green. The third and outer zone begins sharply with a tint of brown which gradually deepens until the outside margin is almost as dark as the core and the birefringence has increased from 0.017 to 0.021.

2. Subhedral to anhedral grains, light green with the color intensity differing in patches or concentric zones without gradation or regularity. The lighter portions have a β index of less than 1.64 and a birefringence of at least 0.028, are optically negative, and seem nearest to tremolite in spite of an extinction angle approaching 25°. The darker ones have a β of at least 1.66, have a lower birefringence and extinction angle, and seem closer to pargasite.

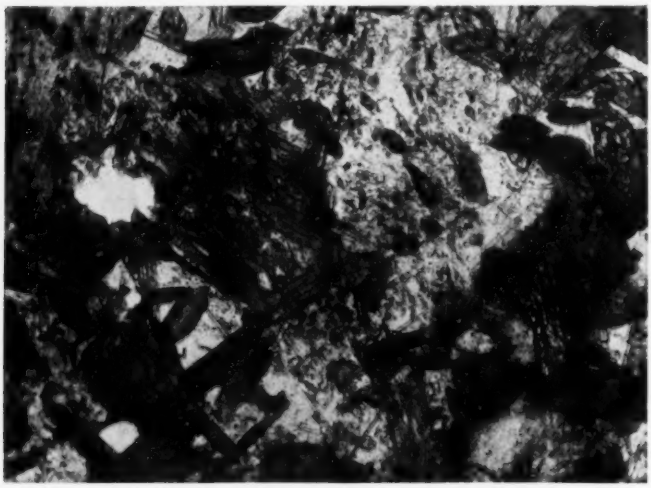

FIG. 2.—Photomicrograph (×100) of the spessartite which was analyzed chemically. Contains idiomorphic amphiboles (darkest) in matrix of saussuritized plagioclase.

3. Microlites near the dike margins differing little, if at all, from the dark core of Type 1.

4. Minute fibers and blades in saussuritized feldspar. "Ears" of apparently similar material extend outward into clear albite from other amphiboles, are pale green with tremolitic birefringence of 0.025 and pargasitic extinction angles up to 30°. The association may be assumed to indicate an aluminous variety.

5. Intergrowths with other minerals of such nature as to suggest replacement of intratelluric phenocrysts. These are gradational into Type 2 above. The intergrowths are of three varieties. One is made up principally of colorless amphibole with scattered granules of

brown mica, clinozoisite and transluscent hematite and with parallel pargasite or actinolite needles forming an uneven marginal fringe. The outlines of the grains suggest their derivation from augite, relict cores of which were found to be common in similar intergrowths in the mica lamprophyre, although no pyroxene was discovered in the spessartite. Another type of intergrowth has the book form of mica; it consists of pale-green tremolite, containing parallel dashes of very fine-grained sphene, and, in most instances, it has cores of calcite. Within the mica lamprophyre is analogous material in less altered form. There flakes of brown biotite are found in all stages of bleaching and transition to tremolite and calcite, with the TiO_2 emerging as rutile needles.

Feldspathic material comprises generally about 40 per cent of the spessartite facies but may be nearly absent in some places, especially near the hornblendite selvages. The lighter areas of the rock, which form a matrix for the amphiboles, consist of saussurite invariably surrounded by clear albite. In the saussurite the constituents identified are, in order of abundance, albite, clinozoisite and zoisite, calcite, actinolite or pargasite, epidote, and sericite. No orthoclase could be found.

Minor constituents of the spessartite include biotite, epidote, apatite, and rutile. Biotite of a brown or brownish-green color occurs in scattered flakes, which are so bunched in some places as to make up as much as 20 per cent of the rock. Epidote, in minute grains, comprises 5 per cent of certain portions; larger grains in the biotite are surrounded by dark-brown, unbleached halos. Apatite is a minor accessory, usually as stout prisms 0.02 mm. thick. It has a birefringence of 0.002. TiO_2 has as its chief manifestation globulitic or nearly microlitic sphene or/and rutile; neither ilmenite nor leucoxene is observed in the spessartite. Magnetite also is absent.

The second facies of the lamprophyre, hornblendite, is found as selvages of a few millimeters in thickness adjacent to the greenstone. It consists almost entirely of microlitic amphibole with more or less fluxional arrangement (Fig. 3), but contains a few minute phenocrysts of the brownish type and a few grains replaced by the fibrous variety. With increase in grain size and the introduction of feldspar the hornblendite grades into the spessartite.

The third facies, mica lamprophyre or malchite, is distinguished

by conspicuous and more abundant biotite. This mineral occurs in 1-mm. plates of light-brown color and shows between its cleavage lamellae networks of rutile needles mingled with red hematite scales. It has a β index of only 1.615—a value which is somewhat lower than that usually found in biotite. In several individual crystals the narrow brown rims form a sharp contrast with the bleached interior. Intergrowths of the amphibole, as mentioned above in the descrip-

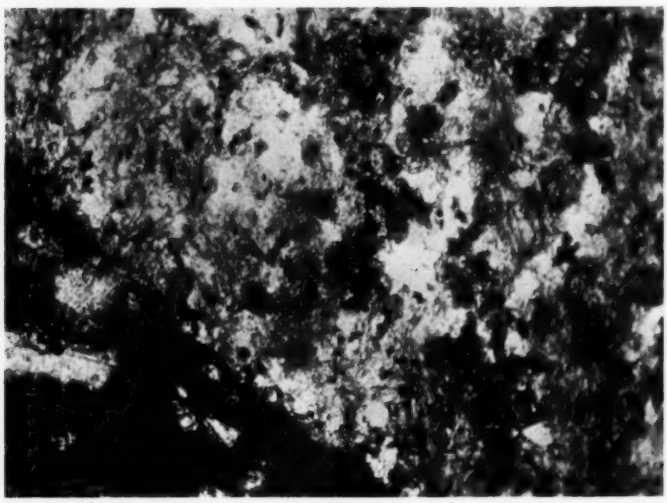

FIG. 3.—Photomicrograph (×100) showing contact of hornblendite (lower left, darker gray) and Mona greenstone schist with irregular, nearly vertical banding.

tion of the spessartite facies, constitute 35 per cent of the mica lamprophyre. One occurrence of a radiated zeolite thought to be natrolite was noted. In other respects this rock resembles the main facies.

CHEMICAL ASPECTS OF THE SPESSARTITE

For quantitative analysis,[5] fragments of the spessartite were chosen which showed under a hand lens no aplite, pegmatite, or vein contamination. Thin sections did reveal vein material, largely calcite,

[5] Performed by L. Gardiner of the Rock Analysis Laboratory at the University of Minnesota.

but in quantities of less than one-quarter of 1 per cent. The rock is compact and unweathered, having had its surface freshened by glacial abrasion. The chemical analysis is given in Table 1.

TABLE 1

CHEMICAL ANALYSIS OF SPESSARTITE

	Per Cent		Per Cent
SiO_2	48.04	TiO_2	0.67
Al_2O_3	13.44	ZrO_2	tr.
Fe_2O_3	3.00	P_2O_5	0.54
FeO	4.38	SO_3	n.d.
MgO	10.33	S	0.24
CaO	9.98	Cr_2O_3	0.06
Na_2O	3.20		99.82
K_2O	1.85	Less O ≡ S	0.09
H_2O+	2.14		99.73
H_2O-	0.11		
CO_2	1.84	Sp. gr.	2.92

TABLE 2

Normative Mineral Molecules	Salic	Femic
or	11.12	.....
ab	27.25	.....
an	16.40	.....
di	.....	14.26
hy	.....	1.96
ol	.....	14.95
mt	.....	4.41
cm	.....	0.09
il	.....	1.22
pr	.....	0.66
ap	.....	1.34
cc	.....	4.20
	54.77	43.09

The C.I.P.W. classification (Camptonose III.I.5.3.4) was obtained from the calculation of the norm as given in Table 2.

For comparison of norm and mode the volume percentages, as

estimated from thin sections, are converted into weight percentages in Table 3.

Attention is called first to the figures for albite. Normal 27.25 and modal 27 are in such agreement that it would seem that all the Na_2O is required for the albite and consequently none is left for the amphibole. Further, since the very definite optical properties of the feldspar eliminate anorthoclase or more potassic members of the series, K_2O does not proxy for Na_2O to any appreciable extent and Na_2O cannot thereby be released from the albite.

TABLE 3

Modal Minerals	Sp. Gr.	Vol. (%)	Wt. (%)
Amphibole	3.07	48	50½
Albite	2.61	30	27
Clinozoisite	3.36	7	8
Biotite	2.95	7	7
Calcite	2.72	4½	4+
Rock	2.92	.....	.....

In the absence of an orthoclasic mineral, the question of the occurrence of K_2O opens up an interesting speculation upon the composition of the average amphibole. Some inferences were attained by deducting from the complete analysis the oxides required to form the other minerals known to be present. The calculations need not be repeated here; they might be misleading because of the approximations required. They tend to indicate an amphibole which differs from ordinary pargasite in being low in lime, high in potash and hydroxyl, and lacking in soda. Attempts at spectrographic confirmation of these peculiarities have not been successful chiefly because of the extreme fineness of the material and the difficulty of obtaining a clean separation of the amphibole from albite. Not one of the spectrographs is free from the sodium lines, and some do not show potassium.

Further comparison of norm and mode of the spessartite serves to emphasize the presence of minerals usually classed as secondary alteration products. Uralite, zoisite, epidote, secondary albite—all are prominently mentioned in descriptions of other rocks with similar

modal specification. For example, the Marquette lamprophyres would be classified by Johannsen[6] as 3112H (or M) close to 2112H. In the latter family is a Minverite from Cornwall with 34 per cent sodaclase (albite), 45 per cent hornblende and augite, 17 per cent epidote and chlorite, 4 per cent calcite, and accessory apatite and pyrite. The rock contains "fresh brown primary hornblende . . . . much brown and purplish augite. . . . . Primary alkali-feldspar, mostly albite, is often abundant. . . . . Often they contain a limited amount of lime-soda feldspar, now replaced by albite, epidote, and prehnite, or in some cases almost wholly albitized."[7] Johannsen comments that "this is probably also a metamorphic rock." In the case of the spessartite, however, we do not believe that the alterations should be considered as metamorphic, because they took place before the introduction of the aplites and pegmatites from the same magma, as we shall amplify later.

PETROGRAPHY OF THE GREENSTONE AND ITS METAMORPHIC PHASES

The banding in the greenstone, ascribed by Williams[8] to bedding in fine-grained pyroclastics, is reflected in differing mineral assemblages and grain sizes. One band which was studied contains grains of clinozoisite and subparallel blades of bluish-green amphibole imbedded in a sparse background of pale chlorite, with leucoxene in scattered clusters of opaque powder. In the adjacent bands the mineral composition is the same but the proportions of amphibole and chlorite are reversed, and much of the clinozoisite has a linear arrangement parallel to the bedding. In a third layer saussurite forms the groundmass (25 per cent), clinozoisite is present as a lacy pattern throughout (40 per cent), the blue-green amphiboles occur in clusters (25 per cent), and brown biotite flakes (5 per cent) and rutile in skeletal and needle forms are characteristic. Other bands are dark colored because of a locally high concentration of amphibole or are striped by podlike streaks of a mineral which resembles sericite.

There is no noticeable alteration in the greenstone either in min-

[6] Albert Johannsen, *A Descriptive Petrography of the Igneous Rocks*, Vol. III (Chicago: University of Chicago Press, 1937), pp. 140–44.

[7] *Ibid.*, p. 142. [8] *Op. cit.*

eral combination or in grain size near its contact with the lamprophyre. Evidences of thermal metamorphism are likewise lacking in the greenstone horses and other inclusions caught within the body of the dike. Such xenoliths are distinguished from the enclosing spessartite by a noteworthy titanium content and by the somewhat bluish tint of the amphibole. The absence of thermal metamorphism at the contact of greenstone and lamprophyre is in striking contrast to effects upon the inclusions of greenstone in the granite referred to later and to those produced by pegmatitic offshoots from the spessartite. Surrounding such offshoots in the laminated greenstone are aureoles composed almost wholly of clinozoisite and epidote.

PETROGRAPHY OF THE GRANITIC XENOLITHS

The rounded granitic xenoliths show considerable variation in composition. Some have almost no ferromagnesian minerals and are either albite granite or albite syenite. Others show streaks or clots of dark-green minerals, chiefly chlorite, in varying amounts; these may be so abundant that the rock might be classified as a quartz diorite gneiss. There can be little doubt that these chloritic streaks represent Mona schist trapped in the border phase of the granitic intrusive.

The dominant felsic mineral in the xenoliths is albite. Andesine also was identified in the gneissic fragments, but it was found to grade into saussuritized feldspar where it comes into contact with the spessartite. Potassic feldspar is subordinate in the granitic rocks, soda-microcline partly replaced by albite being the form in which it was recognized.

A rare but important accessory mineral of the granites is zircon, in slender square prisms of a pale-yellowish-brown color similar to that of the sphene.

The darker bands of the gneiss are essentially recrystallized greenstone. Actinolite, occasionally containing countless minute octahedra of magnetite, predominates over the chlorite in some parts. Enlargement of the accessories (epidote, sphene, and apatite) is outstanding. Such coarsened grains make up a large share of the heavy mineral concentrates from the gneissic granites but are to be found in no other rock of the outcrop. Products of an early Archeozoic vol-

canic activity, they were metamorphosed during the great granitic invasions of the Laurentian and, in later Proterozoic, disrupted and transported by yet another molten upwelling—an unusually interesting geologic history.

CONTACT EFFECTS UPON THE GRANITIC XENOLITHS

Apparently the first effect of the spessartite upon the granitic xenoliths was mere mechanical disruption. The dike rock invaded

FIG. 4.—Photomicrograph (×100) showing the effect of the spessartite upon the granitic xenoliths—insinuating itself into the cracks and prying off quartz grains (white). Two pyrite crystals which were introduced into the granite are shown near the center of the upper third of the picture.

crevices in the xenoliths (Fig. 4) and even pried off fragments of their quartz. High magnifications reveal the initial stage of the rupturing as dots of fluid or pyritic inclusions in rows across the quartz or as tufts of actinolite needles radiating through the quartz from its contact with the dike minerals (Fig. 5). The next minerals to appear were calcite and epidote in narrower cracks. Dislocated fragments of original albite crystals are separated by a comparatively coarse mixture of calcite, epidote, and pyrite. The apophyses in the

xenoliths, which are composed dominantly of albite, are clouded with calcite, epidote, and pyrite (the light gray of Fig. 4), and with subordinate green amphibole.

Surrounding the original granite in some places are narrow zones of the same composition as the apophyses, of which, in fact, they are an extension. These zones furnish the best evidence of reaction between the basic magma and the acidic xenoliths. That they were derived in part from the granite is shown by their isolated quartz

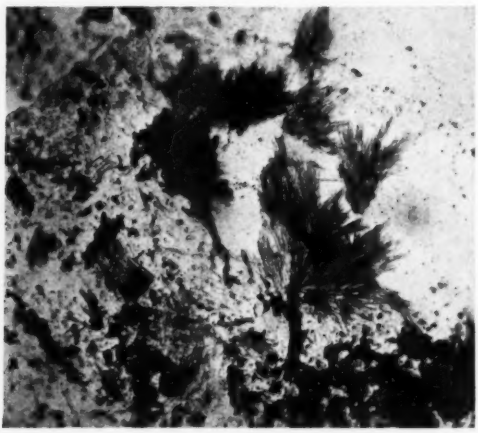

FIG. 5.—Photomicrograph (×440). Effect of spessartite upon the granitic xenoliths—bundles of radiating actinolite needles and rows of fluid or pyritic inclusions penetrate the quartz.

grains, which exhibit more or less complete reaction rims of calcite flakes. The feldspars of these intermediate zones have a cloudiness which distinguishes them from the clear albite of the original granite. Where the two are in contact, the clear variety has a tendency toward cloudy alteration at the margin. The spessartite magma must have furnished, by direct precipitation, a part of the feldspar and nearly all of the lime and iron minerals for the intermediate zones; consequently, the combination is considered as a primary deposition and not as a product of later saussuritization.

Several of the xenoliths were faulted during the later stages of

consolidation of the dike (Fig. 6). The microscope reveals in one of these a 4-mm. mylonitized zone, with gradual lessening in intensity of the cataclastic effects away from the actual break. The fault plane extends outward into the dike, where it is seamed locally with pink pegmatite. The same pegmatitic mineralization is manifest in the mylonite as clinozoisite granules. Fractures in the feldspar of the granite were healed by recrystallization or by fresh growth of albite.

FIG. 6

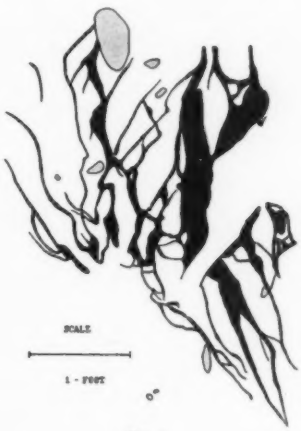

FIG. 7

FIG. 6.—Photograph of 15 in. × 20 in. spessartite surface with faulted xenoliths and aplitic stringers; also showing glacial grooving.

FIG. 7.—Sketch showing details of aplite ramifications (black) in the spessartite (white). Granitic xenoliths shaded.

DIFFERENTIATION PRODUCTS

APLITE

Throughout the spessartite, but more especially in the narrower part of the dikes, occur light-gray or pink stringers and ramifications (Fig. 7) up to a centimeter in width; these are here called "aplites." The gray phase also occurs as thin lenses up to 45 cm. in length, and the pink type as a cement of the crackle-breccia in both the spessartite and the gray aplite. Fragments in the breccia are several centimeters across in the former rock and a few tenths of a

millimeter in the latter. The walls of the pink variety are more sharply defined in the lamprophyre than in the gray aplite.

Microscopically the gray rock (Fig. 8) consists largely of saussuritized and sericitized subhedrons of plagioclase up to 1 mm. in length, with about 25 per cent of clear albite as rims and interstitial filling and with 10 per cent of an amphibole-chlorite-biotite assemblage containing accessory calcite, pyrite, and acicular apatite.

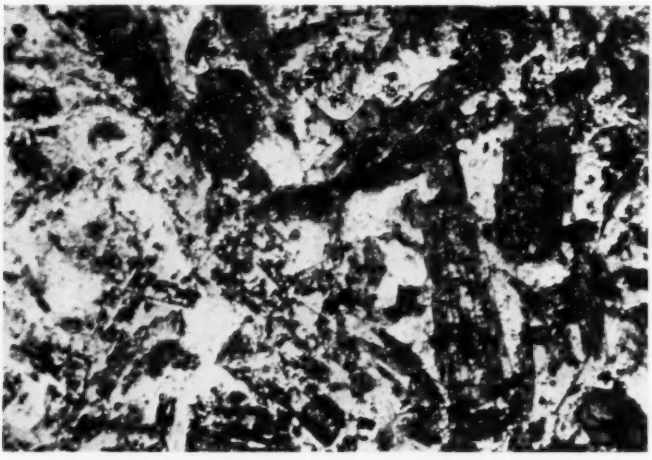

FIG. 8.—Photomicrograph (×100) of an irregular contact extending obliquely from the upper left corner of early aplite (lower left) and spessartite. Color zoning is displayed by some of the larger amphibole prisms.

The pink rock contains less of the saussuritized plagioclase and calcite and more of the clear albite, light-green biotite, and sericite than does the gray phase. Its feldspar grains are smaller, averaging only 0.2 mm., and the amphibole is of a paler-green color. Later stages of the pink aplite contain cores of quartz but no micrographic intergrowths of quartz and orthoclase, such as exist in the diabases of the vicinity.

PEGMATITE

Pegmatite is found in sharply defined pink dikelets. It occurs principally within the lamprophyre, but in places it extends into the crests of small flexures in the marginal laminated greenstone.

In the greenstone the pegmatite is fringed by parallel bands of a clinozoisite-epidote intergrowth, alternating with thin streaks of quartz, which shows undulatory extinction. The pegmatite itself consists of potash feldspar studded with clinozoisite-epidote and clouded by calcite granules, muscovite shreds, and iron stains in an almost submicroscopic crackle pattern. Clear "windows" in each feldspar grain show the grid twinning of (soda-?) microcline, without detectable differences in orientation or refractive indices between the cloudy and the clear portions. The variable character of the epidote was clearly demonstrated by a (010) section of a phantom crystal whose shells alternated sharply between epidote and clinozoisite. Birefringence and extinction angles indicated a 35–40 and a 5 per cent content of the ferric minal, respectively, in the separate parts.

Pegmatite of a different type occurs in the lamprophyre itself. It has an outer selvage of clinozoisite, fingering into a light-green band of biotite and perhaps of chlorite. The biotite is in recognizable blades, but the greater part of the band is exceedingly fine grained, with antigorite-like structure. Next is a zone of muscovite, in plates which become coarser and which lose their suggestion of greenish-yellow tint toward the center of the stringer. The core of the pegmatite consists of a microscopic intergrowth of muscovite and microcline (Fig. 9) with 2–5 per cent of pyrite and 20 per cent of clinozoisite.

In contrast with these occurrences the pegmatite appears in part as a mere alteration of the minerals which it traverses. An example is shown in Figure 10, where two bleached belts cross an amphibole phenocryst (just to the right of the lower margin of the picture) and continue upward to the left as pegmatite (white in the photomicrograph), intersecting gray aplite (speckled light gray). Here the pegmatite consists of a mosaic of microcline-microperthite in grains up to 0.15 mm., with a smaller amount of albite and quartz. In view of the apparent introduction of potash into the albite aplite by the pegmatite, the question may be asked whether the bleaching of the pargasite is not due to a similar process, especially when the potash content of the modal amphibole is recalled. Elsewhere, particularly in a uralitic paramorph which is salted with epidote granules, the same pegmatitic streak consists mainly of calcite flakes with clusters

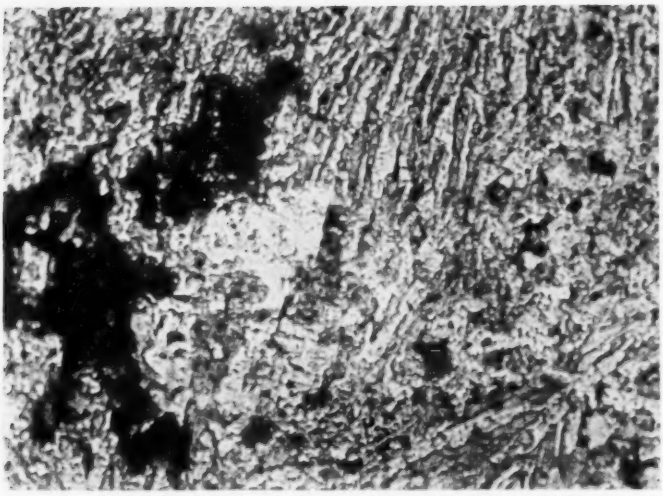

FIG. 9.—Photomicrograph (×440) of a micropegmatitic intergrowth of muscovite (lightest) and microcline (light gray), studded with clinozoisite granules (dark gray).

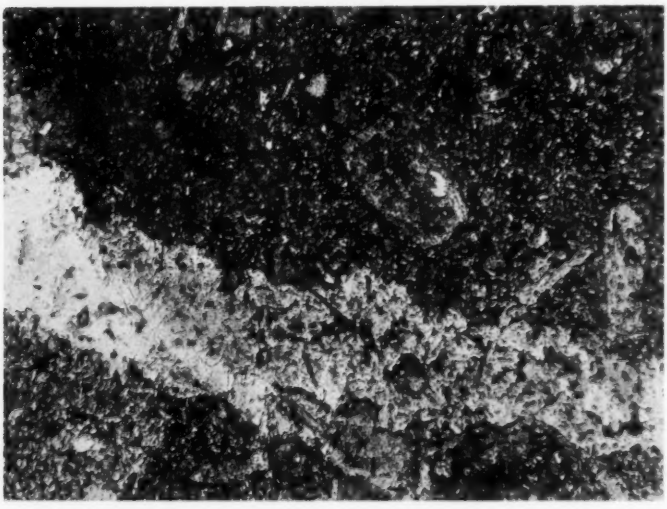

FIG. 10.—Photomicrograph (×30). Spessartite (darker gray) with altered phenocrysts is traversed by aplite (light gray). Pegmatitic replacement crosses phenocryst at center of lower margin and extends upward to the left as a narrow white band across the aplite.

of clinozoisite grains, but its bleaching effect is repeated where it passes through a band of pargasite needles of darker shade.

VEINS

Thin pegmatite seams grade into thinner epidote veins. One of these occupies a fault plane, which transects and offsets the middle and shortest of the three lamprophyre dikes of the outcrop. Coarser veins cut N. 60° E. across the main dike and, passing into the massive greenstone, die out in shatter zones. They are margined by comb epidote and filled by calcite.

The connection of the epidote-calcite veins with the spessartites is assumed partly on the basis of their distribution, which is limited to the immediate vicinity of the dikes, and partly because of their position at the end of the mineral sequence disclosed by this study. These veins are assumed to have no connection with similar veins closely associated with, and confined to, younger basic intrusives of the district.

SUMMARY OF THE COURSE OF CRYSTALLIZATION

When the dikes were emplaced, pyroxene, more or less altered, was probably carried along as suspended crystals in the magma, as was also biotite. Present, too, were fragments of granitic rocks ripped from the wall rocks at some depth. The liquid phase was comparatively rich in hyperfusibles, chiefly water and carbon dioxide, with some hydrogen sulphide. The first mineral to crystallize in the dikes was amphibole, as shown by the chilled border facies of hornblendite. The liquid reacted with the ferromagnesian autocrysts to produce uralite and pargasite, with simultaneous crystallization of more amphibole, largely as phenocrysts. Meanwhile plagioclase had begun to precipitate. The early feldspar, if the equal volumes of zoisite and albite in the saussurite furnish a reliable criterion, must have been an acid labradorite of about Ab_4An_5. The combined plagioclase of the norm approximates Ab_3An_1; from such a liquid the first crystals to be expected, according to Bowen's temperature concentration diagram,[9] would be a basic labradorite, Ab_3An_5. Eventu-

[9] N. L. Bowen, *The Evolution of the Igneous Rocks* (Princeton: Princeton University Press, 1928), p. 34.

ally the early plagioclase, whatever its composition, was made over into saussurite. How quickly this change occurred, or whether the zoisite-albite combination was due to primary deposition from a water-rich magma, is uncertain. In any event, the development of the zoisite-albite must have preceded the complete consolidation of the rock, since the definitely younger, clear albite makes up one-fifth to one-sixth of its bulk. For the main body of the dike the consolidation was completed with the crystallization of the clear albite, but in the openings, owing to shrinkage and shearing cracks, the later stages of consolidation had a separate and independent history. There a freer circulation was maintained and residual liquors accumulated. The liquid of the first cracks gave rise to aplites with only 10 per cent of ferromagnesian minerals (compared with 55 per cent in the spessartite itself) and with much plagioclase containing enough lime to permit light saussuritization. Successively later cracks held liquid which precipitated less and less of these materials and more and more clear albite. Ultimately, muscovite, microcline, and quartz were deposited in very narrow openings, while potash was substituted for soda along the walls of cracks with infinitesimal openings; the pegmatite stage had thus been reached. Where the pegmatites extended out into the greenstone the wall rock was altered to clinozoisite and epidote, with or without actinolite. Epidote also developed as seams in the spessartite along the extension of many of the cracks into which pegmatite had been injected. The depositing solutions were no longer molten—they were approaching the aqueous. Their final products were epidote-calcite veins.

The formation of pyrite was probably begun in the deuteric stage and continued into the pegmatitic. Calcite permeates the products of the deuteric period and may have been formed at the expense of calcium from the amphibole. Calcite dropped to a minimum in the late aplites and pegmatites but reached a maximum in the final veins. Clinozoisite and epidote were contemporaneous with the calcite but reached their greatest development at different times—the clinozoisite during the shearing at the beginning of the pegmatitic stage and the epidote in the later veins.

CONCLUSIONS

Conclusions reached in the study here presented are:

1. Granitic xenoliths occurring in lamprophyric dikes near Marquette furnish not only a means of identifying the members of the dike-swarm but also a possible correlation of their intrusion with the sedimentary record.

2. Three of these dikes studied in detail are spessartites with thin marginal selvages of hornblendite; of these, one grades into malchite.

3. The dike rock was not entirely liquid at the time of emplacement but contained, besides the granitic xenoliths, more or less altered autocrysts of augite and biotite and perhaps of olivine.

4. The marginal chilling of the molten rock produced an extremely fine-grained hornblendite but was not sufficiently rapid to form glass.

5. Repeated cracking in the solidifying spessartite afforded lodgment for the portions then liquid; as a result, there has been preserved a record of the differentiating of the magma by fractional crystallization through aplitic and pegmatitic stages.

6. The average amphibole of the spessartite seems to be high in potash, low in lime, and lacking in soda, and perhaps has an overabundance of hydroxyl.

7. The early formed plagioclase of the spessartite was probably as basic as labradorite but was saussuritized before the final consolidation of the rock.

8. Like the saussurite, much of the calcite was developed in the deuteric stage, and consequently the question of its "primary" character is a matter of definition.

9. Soda was replaced by potash in the pegmatitic stage of crystallization of the lamprophyre.

10. The granitic xenoliths did not fuse in the more basic lamprophyric magma but, instead, reacted with the more volatile constituents to form pyrite, calcite, epidote, saussurite, albite, and pargasite.

11. The physical conditions characterizing the later stages of consolidation of the spessartite involved low temperature and much

water. This conclusion agrees with the views of Barth, Eskola, and Laitakari, as quoted by Johannsen.[10]

ACKNOWLEDGMENTS.—The contributions of the junior author consisted of participation in the mapping, the collecting of samples, and the preparation and preliminary study of thin sections. The writers gratefully acknowledge the aid given by the Michigan College of Mining and Technology, which provided funds for the rock analysis, and the assistance of Reynolds M. Denning, a student, in the preparation of the photographs, and of Drs. R. M. Dickey and Ella Wood, of the Michigan College of Mining and Technology, in valuable criticism of the manuscript.

[10] *Op. cit.*, pp. 142–43.

PROBABLE CONNECTION OF THE NASHVILLE AND OZARK DOMES BY A COMPLEMENTARY ARCH

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ABSTRACT

After the final deformation of the sediments accumulated in a geosyncline, four parallel, genetically associated tectonic belts are believed to occur: (1) deformed hinterland, (2) intensely faulted and folded anticlinorial mountainous belt, (3) folded synclinorium, and (4) a complementary arch which consists of a succession of domelike swells connected by arches or saddles. The occurrence, relationships, origin, and distribution of complementary arches in eastern United States are discussed. Based upon several lines of evidence, a complementary arch is postulated to have connected the Nashville and Ozark domes before it sagged and was covered by sediments of the Mississippi embayment.

INTRODUCTION

For several years the writer has been greatly interested in the arches or flexures that apparently coincided with the foreland rims of Paleozoic geosynclines in the eastern half of the United States (Fig. 1). These include the Appalachian and Ouachita-Llanorian-Marathon geosynclines with their possible connection beneath sediments of the Mississippi embayment and the Arbuckle-Wichita-Amarillo Hills "geosynclines" or alined, and in part en echelon, geosynclines. During the Paleozoic development of these geosynclines three belts, or zones, were probably present: (*a*) the up-arching, sediment-furnishing hinterland or geanticline; (*b*) the downbowing, sediment-receiving geosyncline; and (*c*) the flexure developing along the foreland rim of the geosyncline and characterized by a tendency to stand higher than the geosynclinal belt on one side and the general level of the foreland area on the other side. Along this flexure, or arch, domes, or swells, were the most elevated portions.

These three belts were in existence from very early Paleozoic times; and the relative movements of the geanticlines strongly upward, the geosynclines strongly downward, and the flexures mildly upward occurred at many intervals during the Paleozoic era. During this era the geosynclinal belts received great thicknesses of sediments. Near the close of the Paleozoic, orogenic stresses were applied at various intervals of time westward from the Appalachian

region, northeastward from the Arbuckle-Wichita-Amarillo Hills region, and northward (in general) from the Ouachita-Llanorian

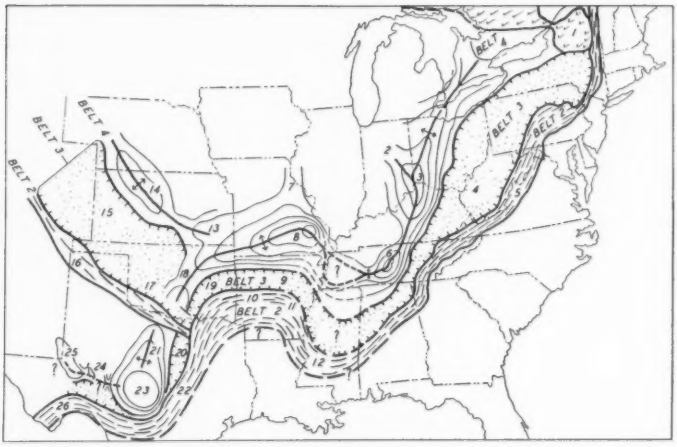

FIG. 1.—Map of eastern United States showing the distribution of genetically associated belts, as follows: intensely folded and faulted anticlinoria (belt 2), indicated by dashed lines; folded synclinoria (belt 3), indicated by stippling; and complementary arches (belt 4), indicated by generalized structure-contour lines. The deformed hinterland (belt 1) is omitted. Pre-Cambrian rocks in the northeastern part of the map are indicated by checks.

1. Adirondack Mountains (swell)
2. Kankakee arch
3. Jessamine swell on Cincinnati arch
4. Folded Appalachian synclinorium
5. Intensely folded and faulted Appalachian Mountains
6. Nashville swell on Cincinnati arch
7. Mississippi River arch
8. Ozark swell
9. Arkansas Valley basin (synclinorium)
10. Ouachita Mountains (folded and faulted anticlinorium)
11. Little Rock salient of Ouachita Mountains, after Miser
12. Postulated connection between the Ouachita and the Appalachian mountains, after Miser
13. Chautauqua arch
14. Central Kansas uplift (swell)
15. Dodge City basin
16. Wichita-Amarillo Hills trend. Arbuckle Mountains are indicated by short line north of eastern end of this belt.
17. Anadarko basin (synclinorium)
18. Central Oklahoma arch
19. McAlester basin (synclinorium)
20. Strawn basin
21. Bend arch
22. Postulated connection between the Ouachita and the Marathon mountains, after several authors
23. Llano uplift (swell)
24. Reagan structure
25. Pecos "uplift"
26. Marathon Mountains

region. These stresses resulted in deformation of the bordering hinterland areas of Appalachia and Llanoria and in the deformation of the great thicknesses of sediments that had accumulated in the

geosynclines. The sediments that had been deposited on the hinterland slope of the geosyncline were on the side of greatest intensity of deformative stresses and were intensely deformed, being uplifted as a deformed anticlinorial belt, whereas the sediments that had been deposited on the foreland slope were farther from the greatest thrusting and hence were less intensely deformed, being bowed downward as a synclinorial belt. The flexure along the foreland rim, with its inherited tendency to rise, was uplifted, although on a much smaller scale.

Following this final deformation of sediments in a geosyncline, four parallel, genetically associated tectonic belts are apparent: (1) the deformed hinterland that is commonly obscured by a subsequent sedimentary covering; (2) the intensely folded and faulted belt corresponding to the hinterland slope of the geosyncline, where in general a highly complex anticlinorium modified by varying numbers of close folds and thrust faults was formed; (3) the foreland slope of the geosyncline, where in general a moderately folded synclinorium was formed; and (4) the accentuated flexure, or "complementary arch,"[1] characterized by an alternation of low saddles connecting domelike swells. In the first three belts deformation was accomplished primarily by horizontal compressive stresses, while in the fourth belt deformation was accomplished primarily by vertically acting stresses, which were in part indirect products of direct horizontal stresses operative in the hinterland and transmitted through the geosynclinal fill and the underlying rocks of the basement complex. The line of separation between belts 2 and 3 corresponds closely to the line of maximum thickness of sediments deposited in the geosyncline[2] and, as a rule, is a distinct, easily recognizable line. The line of separation between belts 3 and 4, however, is an arbitrary line, breaking a continuous dip from belt 4 into belt 3. Three of these tectonic belts are presented diagrammatically in Figures 1 and 2, belt 1 being omitted from these figures.

Many complicating factors enter into such broad structural gen-

[1] This term was used by John L. Rich when reading a paper before the 1936 meeting of the Geological Society of America.

[2] Paul H. Price, "The Appalachian Structural Front," *Jour. Geol.*, Vol. XXXIX (1931), pp. 25 and 43–44.

eralizations; but it is believed that, when viewed diagrammatically, these belts are definite, homologous structural units that can be correlated on the bases of similarity in history during development and deformation and of similarity in resulting structural types. It must be remembered, however, that many sections of these structural belts were developed at different times from those at which other sections of the same belt were developed. The different dates of development of different sections united in one belt in Figures 1 and 2 are not believed to nullify, to any extent, their union by correlation on the two bases given above. Complications due to the time element are more confusing in case of the complementary arches (belt 4) than in case of belts 2 and 3.

Corollary to the fact that the axis of a geosyncline shifts during its development,[3] the complementary arch, being represented during geosynclinal development by the border or rim of the geosyncline, should likewise shift in response to the change in position of the axis of maximum downbowing of the geosyncline for any unit of time. This would be in response to change in locus of the line of maximum depression and deposition and to change in variations of intensity of stresses applied during the same unit of time. The geanticlinal up-arch, geosynclinal downfold, and the foreland arch are all natural mechanical responses of the crust to horizontal stress; and, if the geanticlinal up-arch and geosynclinal downfold move forward (and up and down, respectively), the foreland arch will necessarily move forward (and up).[4] Such shifts in position of the crest of a complementary arch have been referred to by several writers and are well shown for the Chautauqua arch in southeastern Kansas and the central Oklahoma arch in east-central Oklahoma by comparing the published isopach maps, structure contour maps, and paleogeologic maps of these arches for different intervals of time.[5] The same type

[3] M. G. Cheney, "History of the Carboniferous Sediments of the Mid-continent Oil Field," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XIII (1929), pp. 564–81.

[4] W. T. Thom, Jr., personal communication.

[5] Cheney, *op. cit.*, Fig. 7, p. 583; Hugh W. McClellan, "Subsurface Distribution of Pre-Mississippian Rocks of Kansas and Oklahoma," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XIV (1930), Fig. 2, pp. 1542–43; N. W. Bass, "Origin of Bartlesville Shoestring Sands, Greenwood and Butler Counties, Kansas," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XVIII (1934), Fig. 12, p. 1334; T. C. Hiestand, "Regional Investigations, Oklahoma and

of shifting has been shown to have occurred during development of the Nashville dome.[6] Therefore, it is believed that, if the complementary arch is represented not by a single axis but rather by a "composite axis" composed of a relatively wide belt within which the complementary arch stood higher at different loci for different time units, then time relationship is not an unsurmountable obstacle for such structural generalizations as presented in Figure 1.

Also, in addition to complementary arches forming a fourth tectonic belt genetically connected with, and a part of, the system composed of intensely folded and faulted anticlinorial belt 2 and folded synclinorial belt 3, the complementary arches seem likewise to be an integral part of another system of structures within the interior of the continent. This system includes the Kankakee arch, the Mississippi River arch, and the Wisconsin swell, with the intervening synclinal basins.[7] Keith[8] has illustrated this complicating factor of space relationship, showing the way the complementary arch appears to be united with geosynclinal systems, on one hand, and with the pattern of swells, arches, and basins within the continental interior, on the other. His interpretation is that the pattern of cross folds was formed by two sets of cross stresses, one pushing from the Appalachian region on the southeast and the other from the Ouachita region on the south, transmitted by rigid pre-Cambrian rocks

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Kansas," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XIX (1935), Fig. 6, p. 957; Fig. 7, p. 958; and Fig. 8, p. 960; Edward A. Koester, "Geology of Central Kansas Uplift," *Bull Amer. Assoc. Petr. Geol.*, Vol. XIX (1935), pp. 1410–26.

[6] C. W. Wilson, Jr., "The Pre-Chattanooga Development of the Nashville Dome," *Jour. Geol.*, Vol. XLIII (1935), pp. 469–70.

[7] Nemaha Ridge and La Salle anticline are omitted from this group because, even though they have histories essentially parallel to some of the units of complementary arches, they are believed to have originated from, and been located by, buried lines of weakness, as faults, in the underlying rocks of the basement complex (A. E. Fath, "The Origin of the Faults, Anticlines, and Buried 'Granite Ridge' of the Northern Part of the Mid-continent Oil and Gas Field," *U.S. Geol. Surv. Prof. Paper 128-C* [1920], p. 83; Gilbert H. Cady, "The Structure of the La Salle Anticline," *Ill. State Geol. Surv. Bull. 36* [1920], p. 179) and hence are not members of the pattern of tectonic belts, the relative positions of which may possibly have been determined largely by major geosynclinal belts and the history of which has such an intimate relation to that of the major geosynclinal belts.

[8] Arthur Keith, "Outlines of Appalachian Structure," *Bull. Geol. Soc. Amer.*, Vol. XXXIV (1923), Pl. IV, opp. p. 309.

and the more massive Paleozoic sedimentary units.[9] Such a pattern is similar to a system of interference waves, developed in sympathy with the two sets of cross stresses. A section across either of the sets would have intersected an up-arching geanticline (belt 1), a down-bowing geosyncline (belts 2 and 3), a complementary arch (belt 4) which would be a primary wave, a basin (as the Illinois, Michigan, Forest City, and Salina basins), and another arch, or secondary wave, such as the Kankakee and Mississippi River arches.

A homologous succession of "waves," composed of the Yellowstone Park region, Bighorn basin, Bighorn Mountains, Powder River basin, and the Black Hills, was interpreted by Thom, as follows:

> Given an initial elevation in the Yellowstone Park region induced by compressive forces directed eastward, there needs must have been a contemporaneous depression of a trough (Bighorn Basin) to the east under the oblique downward thrust of the uplifted mass. The tangential forces producing the Yellowstone Park uplift would to some extent have affected areas farther to the east. The elastic bending stresses caused by the weight of the newly formed upfold, though progressively diminished outward from the fold by the partial plasticity of the earth's crust, would also help to produce a series of structural waves. . . . .[10]

This series of crests, once established, would tend to continue to rise with reference to the adjacent synclines as a result of several causes cited by Thom.[11]

Bucher, however, believes[12] that the pattern of swells and basins and the processes that cause swells and basins are distinct from those that determine geosynclines and may antedate the initiation of Paleozoic geosynclines. He bases this belief on the fact that swells and basins occur in much the same way on the floor of the Atlantic Ocean and in western Africa. He suggests that those units of the swell-and-basin pattern that occur nearest the independently superimposed Paleozoic geosynclines would be accentuated by orogenic activity in the geosynclines and their hinterlands more than those units farther away from the mobile belt and would hence be welded together in a composite element, the complementary arch.

[9] *Ibid.*, pp. 325–30.

[10] "The Relation of Deep-seated Faults to the Surface Structural Features of Central Montana," *Bull. Amer. Assoc. Petr. Geol.*, Vol. VII (1923), p. 5.

[11] *Ibid.*, pp. 5–6.

[12] W. H. Bucher, personal communication.

Another complicating factor in generalizing on broad regional features in eastern United States is the apparent juxtaposition of a portion of three tectonic belts of the Appalachian-Ouachita-Llanorian-Marathon system with three tectonic belts of the Arbuckle-Wichita-Amarillo Hills system. The junctions of the three tectonic belts of these two systems illustrate *concentric linkage*, as interpreted in Figure 1. Linkage of two complementary arches, of which the loci of greatest "uplift" or minimum subsidence shifted several times during their development, would result in a broad arch, such as the Chautauqua arch in southeastern Kansas, expressing activity in both geosynclinal systems.

ORIGIN OF COMPLEMENTARY ARCHES

Complementary arches, which, during development, coincided with the foreland border of associated geosynclines, were apparently formed by (1) bending downward on the geosynclinal side under the great thickness of sediments deposited on that side, followed by regional tilting in the opposite direction;[13] (2) differential regional subsidence that was relatively great on the geosynclinal side of the arch, or flexure, but relatively slight on the foreland, or continental, side of the arch, with the "axis" of the arch for any unit of time undergoing minimum subsidence within the region;[14] (3) possibly deep subsurface transfer of material from under the subsiding geosynclinal block, resulting in upward components of plastic yielding in the deeper rocks expressing themselves beyond belts 2 and 3, which are dominated by surficial horizontal compressive stresses; (4) possibly an isostatic "upward" movement or positive stand of less dense granitic units of the crust (as the swells occurring along a complementary arch), in contrast to settling of adjacent heavy basic units;[15] and (5) possibly some vertical uplift from vertical components of horizontal compressive stresses transmitted from the hinterland region in pre-Cambrian rocks of the basement complex.[16] Whatever combination of the foregoing factors predominated, differential

[13] A. I. Levorsen, "Convergence Studies in the Mid-continent Region," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XI (1927), Fig. 15, p. 682; Cheney, *op. cit.*, p. 573.

[14] Cheney, *op. cit.*, p. 574; Wilson, *op. cit.*, p. 471.

[15] F. B. Plummer and R. C. Moore, "Stratigraphy of the Pennsylvanian Formations of North-Central Texas," *Univ. Texas Bull. 2132* (1921), pp. 201–2.

[16] Keith, *op. cit.*, p. 341.

uplift or less subsidence was at a maximum in those parts of the arch occupied by swells and at a minimum in the connecting saddles.

The distance of a swell from the axis of the adjacent geosyncline (and hence from the subsequent intensely deformed belt 2) may be controlled by several factors. If a swell owes its character to the presence of less dense, granitic rocks, its location is predetermined by the area of such rocks, and its tectonic history as a positive element will show reflections of activity in the nearest geosyncline (or geosynclines, should two be moderately close). In that case the width of synclinorial belt 3, which separates the swell from the more intensely deformed part of the geosyncline (belt 2), should correspond to the width of the negative area of heavy, basic rocks with its tendency to downward movement inherited from the early tectonic history of the region. Such a downbowing belt would act as a locally effective barrier to upward expression of lower plastic yielding during both developmental and deformational stages. It is also possible, however, that the pattern of swells and basins, and also their tendency to continue to act as positive and negative elements, respectively, are entirely independent of the distribution of granite and metamorphic rocks in the basement complex and instead were inherited from structural events of pre-Cambrian tectonic history.[17] By one or the other, or both, of these conditions the writer would explain the apparent anomalies, (1) that the central Kansas uplift (swell) is separated from the Arbuckle-Wichita-Amarillo Hills belt 2 by an abnormally wide synclinorial belt 3 (Anadarko and Dodge City basins), and (2) that synclinorial belts 3 are abnormally narrow, or locally not expressed, in the cases of the Llano uplift and the Adirondack Mountains (swells). The Llano swell is close to, or practically adjoining, the intensely faulted and folded belt of the buried connection between the Ouachita and the Marathon Mountains as it swings around the southeastern part of the swell, and the Adirondack swell practically adjoins the intensely faulted and folded belt 2 of the Appalachian Mountains.

"SOUTHWESTWARD CONTINUATION" OF THE NASHVILLE DOME

Several writers have postulated a southwestward continuation of the axis of the Nashville dome into Alabama and Mississippi.

[17] Bucher, personal communication.

Jillson[18] and Bailey,[19] for example, show the axis of the dome continuing southwestward across northwestern Alabama, through the Amory gas field, Mississippi, and extending toward Jackson, Mississippi. More recent studies,[20] however, show the main axis of the Nashville dome swinging southwestward across the southern part of Maury County, Tennessee, to the east-central part of Wayne County, Tennessee, where a major split occurs, one branch continuing westward across Hardin County, Tennessee, whereas the second branch turns northwestward as far as the central part of Decatur County, Tennessee. In the central part of Hardin County and in the central part of Decatur County the respective branches disappear beneath the Cretaceous sediments of the Mississippi embayment, the former axis heading westward, the latter northwestward (Fig. 2).

"CUMBERLAND RIVER ARCH"

The position of this arch was indicated by Jillson[21] in 1931 as crossing western Kentucky in a northwest-southeast direction parallel to the Tennessee and Cumberland rivers. It is doubtful if such an arch exists in the position indicated by Jillson. Weller,[22] in a structure-contour map of Illinois and parts of adjacent states, shows only a northeast dip off the Ozark dome in that particular part of Kentucky. Rhoades, who has studied the geology of the lower part of the Tennessee River and the Cumberland River drainage basins, reports[23] that he has been unable to find evidence for the postulated "Cumberland River Arch" in the position shown by Jillson.

PROBABLE CONNECTION BETWEEN THE NASHVILLE AND OZARK DOMES

It is believed that a low saddle, such as the one now connecting the Jessamine and Nashville domes along the Cincinnati arch, con-

[18] W. R. Jillson, "Geology of Amory, Mississippi, Gas Field," *Oil and Gas Jour.*, Vol. XXVI (1928), p. 58.

[19] W. F. Bailey, "Natural Gas from Paleozoic Horizons in Southern Cincinnati Arch Region," *Geology of Natural Gas* (Tulsa: Amer. Assoc. Petr. Geol., 1935), Fig 1, p. 859.

[20] C. W. Wilson, Jr., and E. L. Spain, Jr., "Upper Paleozoic Development of Nashville Dome, Tennessee," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XX (1936), Fig. 2, p. 1078.

[21] "Structural Geological Map of Kentucky," *Ky. Geol. Surv. Ser. 4* (1931).

[22] J. Marvin Weller, "Geology and Oil Possibilities of the Illinois Basin," *Illinois Petroleum* ("Ill. State Geol. Surv., Press Bull. Ser.," No. 27 [1936]), Pl. I, pp. 10–11.

[23] Roger F. Rhoades, personal communication.

nected the Nashville and Ozark domes before it sagged and was covered with sediments of the Mississippi embayment. Such a saddle is believed to have existed at approximately the same structural elevation as, or possibly somewhat lower than, the saddle between the Jessamine and Nashville domes prior to the beginning of downfolding or downfaulting of the northern end of the Mississippi embayment. This saddle would form the line of dip reversal between the dip into the Illinois basin and the southern extension of this basin into the "low" of Stewart, Houston, Humphreys, Montgomery, Dickson, and Robertson counties, Tennessee,[24] to the northeast and east, respectively (Fig. 2), and the deeper part of the Mississippi embayment to the southwest. The two suggested alternative trends of this hypothetical saddle are shown in Figure 2. The reasons for such a hypothesis are summarized briefly, as follows:

1. The most recent postulation of connection between the Appalachian and the Ouachita Mountains (tectonic belt 2) is by Miser,[25] in which a tentative position of the eastern limb of the Little Rock salient of the Ouachita Mountains is postulated. If this, or some similar connection, is accepted, and if the present writer's presentation of the normal existence of four parallel tectonic belts is accepted, then the belt of intensely faulted and folded rocks (belt 2) connecting the Appalachian Mountains with the Ouachita Mountains should have parallel to it on the north, northeast, and east a similarly curving synclinorial belt 3, connecting the exposed southern end of the Appalachian coal basin with the exposed eastern end of the Arkansas Valley–McAlester basin. Likewise, parallel to belts 2 and 3 there should be a complementary arch connecting the "high" in Wayne County, Tennessee (where the axis of Nashville dome splits), with the southeastern portion of the Ozark dome, as exposed. The two most logical alternative positions for this connecting segment of the complementary arch are shown in Figure 2, the southern one paralleling the Appalachian-Ouachita connection (after Miser) by continuing westward almost to Mississippi River and then swing-

[24] Wilson and Spain, *op. cit.*, Fig. 2, p. 1078.

[25] Hugh D. Miser, "Relation of Ouachita Belt of Paleozoic Rocks to Oil and Gas Fields of Mid-continent Region," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XVIII (1934), Fig. 2, p. 1064.

ing northward, and the northern one cutting northwestward across the embayment. Either of these alternative connecting segments would make a continuous complementary arch from the Canadian Shield to the Llano uplift, and possibly to the Rocky Mountain

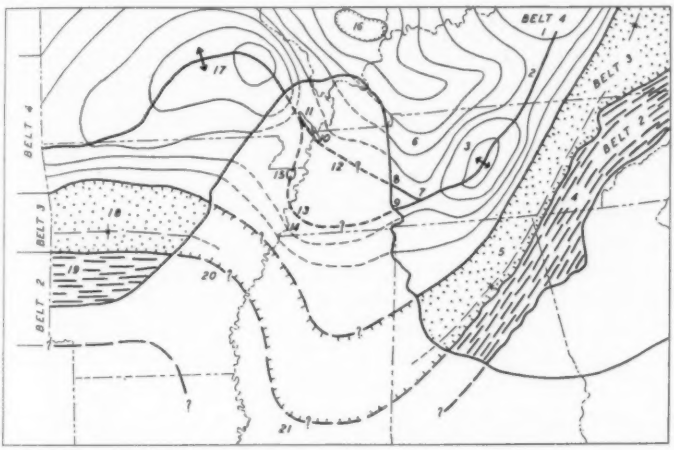

FIG. 2.—Map of the Mississippi embayment and surrounding area, showing connection of tectonic belt 2 (dashed lines) across the embayment, after Miser, and the suggested connections of tectonic belts 3 (stippled) and 4 (generalized structure contour lines), as postulated here.

1. Jessamine swell on Cincinnati arch
2. Saddle on Cincinnati arch between Jessamine and Nashville swells
3. Nashville swell
4. Intensely folded and faulted Appalachian Mountains
5. Folded Appalachian synclinorium
6. Structural "low" occupying parts of Stewart, Houston, Humphreys, Montgomery, Dickson, and Robertson counties, Tennessee
7. Wayne County, Tennessee
8. Decatur County, Tennessee
9. Hardin County, Tennessee
10. Tiptonville "dome"
11. Sikeston ridge
12. Postulated northern arch connecting Nashville and Ozark swells
13. Postulated southern arch connecting these swells
14. Shelby County, Tennessee
15. Blytheville "dome"
16. Illinois basin
17. Ozark swell
18. Arkansas Valley basin (synclinorium)
19. Arkansas Ouachita Mountains (folded and faulted anticlinorium)
20. Little Rock salient of Ouachita Mountains, after Miser
21. Postulated connection between the Ouachita and the Appalachian mountains, after Miser

front, which is broken only where interfered with by linkage with the Arbuckle-Wichita-Amarillo Hills system (belt 2). The Adirondack Mountains would be homologous to a swell along a complementary arch but occupying a position off the periphery of the southeastern

curve of the shield and out of line with the Cincinnati arch. The complementary arch westward from the Llano uplift is obscure, owing to deep burial and marked downbowing of the Permian basin, but is possibly represented by the western slope of Llano uplift, by the Reagan uplift, and by the Pecos uplift,[26] should this "uplift" prove to be of tectonic origin. These uplifts must have been near the foreland rim of the Marathon-Llanorian geosyncline.

2. The many unconformities found in Hardin[27] and Wayne[28] counties, Tennessee, present a difficult problem to anyone trying to compile and to harmonize the Paleozoic history of the western half of Tennessee. Some of these unconformities correspond to breaks occurring in central Tennessee (higher on the Nashville dome), and others occur within groups of rocks that are conformable in central Tennessee, but still other parts of the stratigraphic section in Hardin and Wayne counties are conformable at horizons characterized by breaks in central Tennessee. The Paleozoic history of these counties was in part "out of step" with that of central Tennessee. In 1935 the writer published his conception of synchronism between differential downbowing of a geosyncline and the relative vertical position of an "inland" dome such as the Nashville dome.[29] His conclusions were that, when Appalachia and the associated Appalachian geosyncline were being subjected to tectonic stresses, the Nashville dome was relatively high and unconformities resulted; whereas, when Appalachia and the geosyncline were not being, or had not recently been, subjected to tectonic stresses the dome had relatively low relief and was covered by conformable sediments.

Hardin County is only about 30 per cent closer to the nearest part of the folded and faulted belt 2 of the Appalachian system than it is to the position of the Little Rock salient of the Ouachita Mountains, as interpreted by Miser. Accepting synchronism of movement between a downbowing geosyncline and a complementary arch intro-

[26] E. H. Sellards, "The Pre-Paleozoic and Paleozoic Systems in Texas," *Univ. Texas Bull. 3232* ("The Geology of Texas," Vol. I [1932]), p. 52.

[27] W. B. Jewell, "Geology and Mineral Resources of Hardin County, Tennessee," *Tenn. Div. Geol. Bull. 37* (1931), pp. 20–45.

[28] Miser, "Mineral Resources of the Waynesboro Quadrangle, Tennessee," *Tenn. State Geol. Surv. Bull. 26* (1921), pp. 15–25.

[29] Wilson, *op. cit.*, pp. 464–71.

duces the complication of having the tectonic movements that occurred in the southern end of the Appalachian geosyncline, as exposed, and those in the eastern limb of the Little Rock salient of the Ouachita Mountains, as restored, both influencing the depositional history of the Hardin County and Wayne County "nose" of Nashville dome. As movements in these geosynclines, which were probably connected, did not necessarily coincide, the depositional record in these counties should be broken by more unconformities than that preserved on the flanks of the Nashville dome and by unconformities at horizons characterized by conformity of sediments on the dome. This is believed to be the case.

3. One of the most striking features found by Born in his study of subsurface conditions in the Mississippi embayment region of Tennessee was an east-west structural "high," shown by structure contours on top of the Midway group.[30] This structure extends as far west as Shelby County, Tennessee, through the southern tier of counties just north of the Mississippi state line, and lines up with that branch of the axis of Nashville dome which disappears beneath sediments of the embayment in the central part of Hardin County (Fig. 2). A structural trend in the Midway group suggests rejuvenation along an anticline in Paleozoic rocks beneath sediments of the embayment at some date subsequent to the deposition of the Midway group. Such a Paleozoic anticline would probably be a segment of the postulated saddle connecting the Nashville and Ozark domes.

4. Within the flood plain of the Mississippi River in the general region under discussion there occur several "ridges" that rise from 10 to 25 feet above the level of the flood plain. Four of these—Tiptonville "dome" in northwestern Tennessee, Blytheville and Little River "domes" in northeastern Arkansas, and Sikeston ridge in the southeastern part of Missouri[31]—probably represent upwarped portions of the otherwise graded flood plain. Glenn concludes from his

[30] Kendall E. Born, "Notes on the Upper Cretaceous and Tertiary Sub-surface Stratigraphy of Western Tennessee," *Jour. Tenn. Acad. Sci.*, Vol. X (1935), Fig. 4, p. 261.

[31] M. L. Fuller, "The New Madrid Earthquake," *U.S. Geol. Surv. Bull. 494* (1912), Pl. 1, opp. p. 8.

study of several of these "ridges" that the differences in elevation of alluvial deposits indicate that the region has undergone differential elevation—

> in other words, has been warped . . . . and if earthquake movements be thought inadequate to produce the effects seen, more quiet and slowly acting but more general and powerful crustal movement must be assumed as the cause, for the difference in elevation of the strata . . . . does not seem due to differences in surface erosion.[32]

The Sikeston and Tiptonville "anticlines" are of special interest, since, together, they are 45 miles in length and have a north-northwest strike that is directly across the concentrically curving belts of Cretaceous and Eocene formations exposed in this part of the Mississippi embayment. This relatively long structure cutting across younger structural trends suggests slight, relatively recent surface expression of renewed movement along a buried Paleozoic structure. Such a buried structure might be a weak anticline, a fault, or both.

It is thought significant that these "ridges," which suggest late activity of buried structures, occur along one or the other of the two lines postulated to represent alternative arches between the Nashville and Ozark domes. Since these "ridges" occur along the belt of maximum downwarping of the Mississippi embayment and since the northern two occur in an area broken by subsurface faults, as indicated by the New Madrid earthquake and as suggested by the many faults in existence in the bordering Paleozoic rocks of Kentucky and Illinois, any subsurface structures would be deeply downbowed and probably obscured by major faults. Later downbowing along the north-south axis of the embayment approximately normal to the earlier anticlinal arch should set up local torsional stresses that would cause locally associated faulting.

5. Much geophysical prospecting has been done in the northern part of the Mississippi embayment, but relatively few of the resulting data are now available. Magnetometer maps by Spraragen[33] and

[32] L. C. Glenn, "Underground Waters of Tennessee and Kentucky West of Tennessee River and of an Adjacent Area in Illinois," *U.S. Geol. Surv. Water Supply Paper, No. 164* (1906), p. 19.

[33] L. Spraragen, "Magnetometer Study of Arkansas," *Oil and Gas Jour.*, Vol. XXVII (1929), pp. 30 and 95.

Jenny[34] were used for local control in locating the suggested alternative positions of a complementary arch across the embayment.

CONCLUDING STATEMENT

This paper was written with full realization that no positive proof is now known either for the actual existence of, or for the location of, the postulated saddle connecting the Nashville and Ozark domes. With the scanty details now available the writer merely suggests possibilities and attempts to picture the structure of part of the Mississippi embayment in terms of the tectonic pattern of the eastern United States as he visualizes it.

ACKNOWLEDGMENTS.—Discussion and correspondence with Kendall E. Born, Walter H. Bucher, R. T. Chamberlin, Ira H. Cram, Robert H. Dott, L. C. Glenn, W. B. Jewell, John L. Rich, W. T. Thom, Jr., and T. E. Weirich helped to establish and clarify the principles and details presented here. Scores of published reports were studied in preparation for this brief paper, but to give credit to each author from whose report data were absorbed would make an abnormally long bibliography.

[34] W. P. Jenny, "Oil and Gas Possibilities in Northeastern Arkansas and Western Tennessee," *Oil Weekly*, Vol. LXXXVI (1937), pp. 26–32.

SILURIAN SEA BALLS

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ABSTRACT

The "Niagaran" dolomite in the vicinity of Chicago is characterized by several kinds of nodules, one type consisting of an intimate mixture of silicified fossils in a matrix of calcium carbonate. The mode of origin of these nodules is compared with that of modern "sea balls" and "lake balls," which are produced by wave action in shallow waters. The apparently dwarfed fauna of the Silurian sea balls comprises most of the common Paleozoic classes of invertebrates, with bryozoans, brachiopods, corals, and echinoderms most abundant. An extended storm period has been posulated to account both for the dwarfing of the fauna and for the genesis of the nodules.

INTRODUCTION

Siliceous nodules are common in the "Niagaran" dolomite in most of the many quarries of the "Greater Chicago" area. The nodules may be classified under three generalized categories, as follows: (*a*) irregular to subspherical masses of amorphous chert devoid of organic remains (Fig. 1); (*b*) irregular nodules and lenses apparently formed from gelatinous silica with inclusions of fragmentary fossils which are dominantly echinodermal in origin (Fig. 2); and (*c*) subspherical nodules that consist of numerous silicified fossils embedded in a matrix dominantly of calcium carbonate (Figs. 3 and 4). This latter type of nodule was discovered in 1930 by Professor J Harlen Bretz, of the University of Chicago, and Dr. E. H. Stevens, now of the Colorado School of Mines. Owing to obvious paleontological aspects of the problem of their origin, the first nodules found were presented to the senior author for investigation. Subsequent work has added much material to the collection and has confirmed original opinion as to the importance of the subject.

DESCRIPTION OF THE NODULES

All of the nodules of category *c* are subspherical and some are nearly circular in outline. Most of them, however, have a depressed elliptical section. They range in maximum dimension from about 5 to 20 cm. and are so oriented that their long axes are nearly or quite parallel to the bedding. This suggests that compaction has produced

a flattening of once nearly spherical nodules which were not indurated at the time of deposition. Furthermore, the nodular masses are enclosed by shaly laminae, which are deflected above and below the nodules (Figs. 3 and 4) to produce an undulant stratification. Very fine, nearly vertical slickensides occur along the steeper portions of the flanks of the nodules (Fig. 4). These structural features also indicate a syngenetic relationship between the nodules and the enclosing strata, since the nodules must have been present before

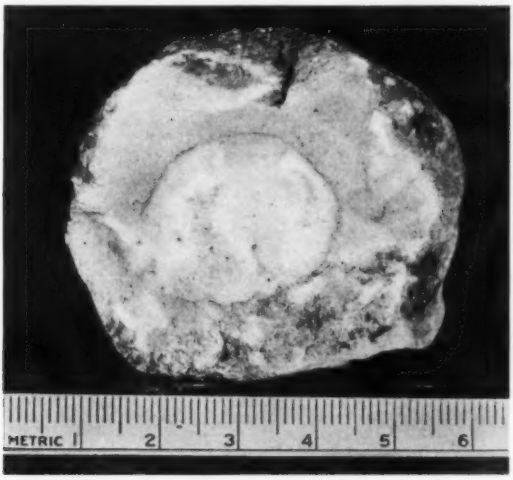

FIG. 1.—Sectional view of a nonfossiliferous chert nodule from quarry at Elmhurst, Illinois.

the rock was indurated. That in most cases they suffered no great compaction probably was due to their strong core of intermeshed fossils.

NODULE GENESIS

Conditions which would be conducive to the formation of such nodular masses may have occurred during periods of storm. Bioherms, which occur near the stratigraphic level of the nodules, suggest relatively shallow Niagaran seas. Occasionally, therefore, great storms might not only have lowered the effective wave base but also

FIG. 2.—Part of an irregular chert nodule with inclusions of fragmental fossils from quarry at Elmhurst, Illinois

might have swept into the sea much more argillaceous material than normally. Thus, turbulent wave action may have become prominent in a normally quiet environment. Such wave action could conceivably have rolled clastic lime muds and argillaceous material on the sea floor into balls, with fossils serving as the nuclei. Presumably such balls would have continued to grow as the rolling action of the

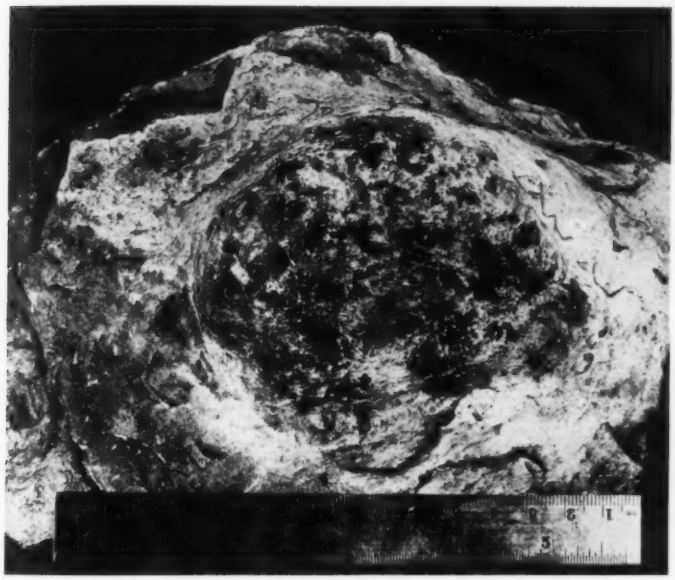

FIG. 3.—View from above showing shaly laminae inflected on flanks of a fossiliferous "sea ball" from Federal Quarry, Chicago.

waves enabled them to pick up more fossils and matrix. Storm duration was doubtless one factor limiting the size of nodule growth and storm violence and depth of water presumably were others, since in part they controlled strength of wave action. During such storms as postulated, coarser material would be carried far from the source of supply owing to the increased strength of the agencies of transportation. Thus, shaly nodular layers in otherwise nearly pure dolomite would be not anomalous but anticipated. Naturally the same

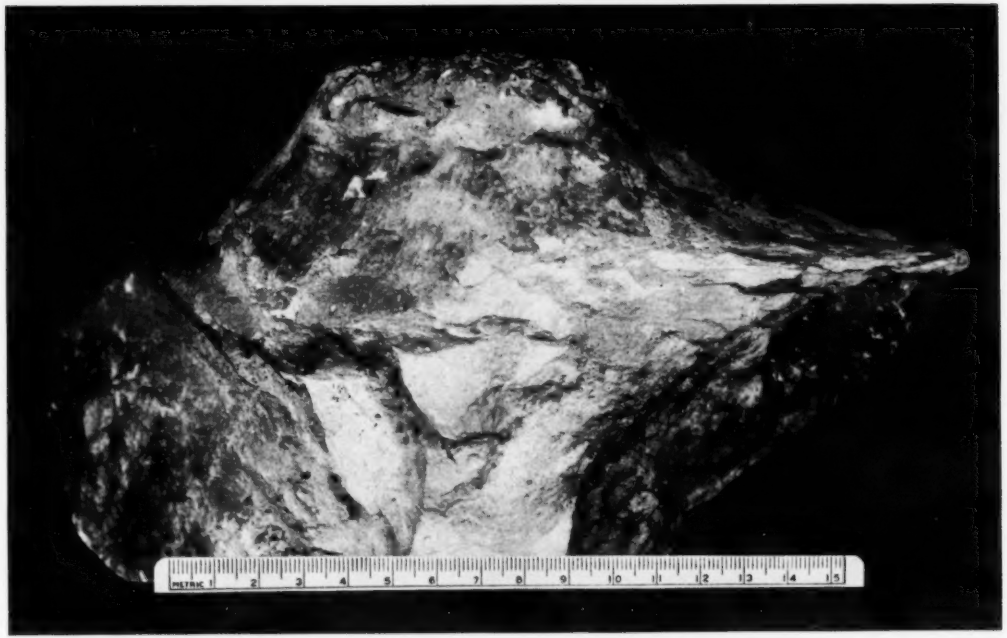

FIG. 4.—Fossiliferous "sea ball" protruding from the enclosing sediments. Slickensides along the left flank of the ball may be seen with the aid of a lens.

general effects might well have been obtained by a temporary lowering of sea level, but a nondiastrophic explanation seems somewhat less strained.

DIAGENETIC SEQUENCE

Dolomitization probably occurred subsequent to induration, since the most favorable conditions for dolomitization presumably involve pressure and long periods of time. The argillaceous materials of the nodules and shaly laminae may have inhibited their recrystallization as dolomite, for they have remained relatively unaltered. Preliminary staining tests show that the nodules and associated laminae have only a few dolomite crystals, whereas the adjacent rock is composed of nearly pure dolomite.

Silicification also probably occurred after induration. Percolating ground waters, carrying a high percentage of silicon dioxide in solution, moved along the shaly zone which may have served both as a conveyor and a concentrator. The nodule matrix and the laminae did not readily exchange materials with the waters, but the fossil shells of calcite and possibly of aragonite were easily removed by solution and replaced by silica. Silicification probably preceded dolomitization; otherwise the fossils would have been partially obliterated or at least markedly injured, owing to the shrinkage phenomena which commonly accompany dolomitization. The almost perfect preservation of some of the fossils appears, therefore, to controvert the suggestion that they were once dolomite. If, however, it is admitted that because of their somewhat argillaceous blanket the fossils might have escaped dolomitization during the diagenesis of the main body of rock, then their silicification may have taken place even as late as relatively recent times.

RELATIONSHIP OF NODULES TO SEA BALLS

Careful study of the fossiliferous nodules, their fauna, and their probable origin has indicated their close relationship to subspherical structures which for centuries have occasionally come to the attention of naturalists. These rounded masses or balls, dominantly of either plant or animal matter, have been found in both fresh and salt water, and there is a scattered, but not very voluminous or definitive literature concerning them. They have been described

under so many different names, however, that references are easily overlooked. Sea balls were known to Moorish scholars, who called them *pilae marinae*. According to Schröder,[1] an Andalusian, Abul Abbas, noted them in his book of travel, *Kitab ar rihlat*, about the year 1216. The early Arabian pharmacologist, Ibu el Baither, in his *Kitab al Dschamie al Kabir*, called these structures *okr elbar* and pointed out their use in medicine, especially in dentistry.

Linnaeus gave the name *Aegagropila* to ball-building cladophores, and Kützing later elevated this name to generic significance. Later still the sea balls came to be called *Aegagropiles de mer*, *Aegagropiles marines*, *Pilotes marines*, or simply *Seebälle*. Thoreau observed similar structures as early as 1845 and described them as lake balls in his *Walden*.[2] Weddel[3] in 1879 first showed in print that some of the pellets were not of algal origin at all, as most observers before him apparently thought, but rather that they were the fringed-out and balled-up fibrovascular bundles of a naiadaceous phanerogam, *Poisedonia caulini*. And gradually, in the last fifty years, evidence has been accumulating to show that the balls may be made up of amazingly diverse materials.[4] McAtee,[5] for instance, has described river balls of salmon bones from the Karluk River, Alaska, and buffalo balls, largely of fragments of *Ruppia maritima*, from a lake near Dawson, North Dakota. Russell[6] has noted balls of fir needles from

[1] Bruno Schröder, "Über Seebälle," *Naturwiss.*, Vol. VIII (1920), pp. 799–803.

[2] Henry D. Thoreau, *Walden* (Boston: Thomas Y. Crowell & Co., 1854), p. 205.

[3] H. A. Weddel, "Sur les Aegagropiles du mer," *Actes du Cong. d. Bot. tenue à Amsterdam au 1877*, pp. 58–62. (Also abstracted in Just's *Jahresberecht*, Vol. IX [1819], p. 333.)

[4] Elizabeth Acton, "On the Structure and Origin of 'Cladophora Balls,'" *New Phytologist*, Vol. XV (1916), p. 9; J. Adams, "Vegetable Balls," *Science*, Vol. XIX (1904), p. 926; A. E. Atwood, "Meetings of the Botanical Branch," *Ottawa Naturalist*, Vol. XVII (1903), p. 41; G. W. W. Barclay, "On Some Algoid Lake-Balls Found in South Uist," *Proc. Roy. Soc. Edinburgh*, Vol. XIII (1885–86), pp. 845–48; W. F. Ganong, "On Balls of Vegetable Matter from Sandy Shores," *Rhodora*, Vol. VII (1905), pp. 41–47; F. R. Kjellman, "Zur Organographie und Systematik der Aegagropilen," *Nova acta reg. Soc. scientiarium Upsaliensis*, Vol. XVII, Fasc. 1, ser. 3 (1908), p. 23; J. R. Lorenz, "Die Stratonomie von *Aegagropila Sauteri*," *Denkschrifter der K. Akad. d. Wissensch., Mat.-Naturw. Klasse*, Vol. X (1855).

[5] W. L. McAtee, "Notes on Drift Vegetable Balls and Aquatic Insects as a Food Product of Inland Waters," *Ecology*, Vol. VI (1925), p. 299.

[6] M. Wm. Russell, "Sur les pelotes marines," *Rev. gén. de bot.*, Vol. V (1893), pp. 65–73.

lakes of the Engadine, and Huntsman[7] and Ganong[8] have obtained similar spill balls, as they are called locally, from Little Kedron Lake, New Brunswick. Balls made largely from tamarack leaves have been reported from a lake in Oakland County, Michigan, and hair balls up to 5 inches in diameter were relatively common on the beach of Little Traverse Bay, near Petoskey, Michigan, up to 1930. At about that date a tannery, which had been dumping hair, tanbark, and other refuse into the bay for more than a score of years,

FIG. 5.—Recent "hair balls" from beach of Little Traverse Bay, near Petoskey, Michigan.

was forced to make other disposal arrangements. Thereafter the hair balls have become increasingly uncommon until they are now rare. Some of the Petoskey hair balls started as long rodlike structures. The pellets become more nearly spherical with increase in size, as is suggested in Figure 5, which shows two of these balls.

Mackay[9] reported that weed balls were common on the beach at Upper Kingsburg, Lunenburg County, Nova Scotia, in February, 1906. Prior to this date, however, they had not been noted by the fishermen. MacKay's informant, a local schoolteacher, counted 200

[7] A. G. Huntsman, "On the Formation of Lake Balls," *Science*, Vol. LXXXII, No. 2122 (1935), pp. 191–92.

[8] "On Vegetable or Burr Balls from Little Kedron Lake," *Bull. Nat. Hist. Soc. New Brunswick 23*, Vol. V ([1904], 1907), pp. 304–6.

[9] A. H. MacKay, "Water-rolled Weed Balls," *Proc. and Trans. Nova Scotia Inst. Sci.*, Vol. XI ([1906], 1908), pp. 667–70.

weed balls in a short space on the beach. The spheres ranged from $1\frac{1}{2}$ to 5 inches in diameter, and there were as many as "400 minute shells from the center of one ball." This latter point may be of some significance in the present study. If seaweeds were responsible for the origin of the Niagaran nodules, however, they have left no definite evidence of their former presence.

ORIGIN OF SEA BALLS

There have been many different opinions as to the mode of origin of sea balls. Lorenz,[10] as early as 1855, considered them to have had a biological-mechanical formation. Ganong[11] maintained that all the balls have one thing in common, i.e., their formation depends upon the rolling action of submerged portions of waves working upon fibrous substances resting lightly upon sandy bottoms. Huntsman[12] has agreed that this indeed may have been the mode of origin of balls that show concentric lamination, but he believes that those which have a radial, rather than a laminar, structure must have achieved their rounded outlines by being worn away by rolling on the bottom or on the beach itself. Kindle[13] and others have pointed out that *Cladophora* balls do result from the radial growth of this alga: and, according to Schröder,[14] there are a number of living marine animals and plants which may take on spherical shapes when they are rotated by wave action. Kindle also has shown that sea balls may be formed in shallow water by a temporary increase of the vigor of wave action.[15] He observed, on the beach a few miles south of Atlantic City, New Jersey, many rounded clay balls up to 10 inches in diameter, which had formed as the result of intense wave action accompanying the hurricane of mid-September, 1936. The balls were firmly bound together by salt grass roots and had many

[10] "Ergänzungen zur Bildungsgeschichte der sog. 'Seeknödel' (Aegagropila Sauteri Kütz.)," *Verhandl. d. K. K. Zool.-Bot. Gesellschaft in Wein*, Vol. LI (1901), pp. 364–65.

[11] "On Balls of Vegetable Matter from Sandy Shores," *op. cit.*, Vol. XI (1909), pp. 149–52.

[12] *Op. cit.*

[13] E. M. Kindle, "Concerning 'Lake Balls,' 'Cladophora Balls,' and 'Coal Balls,'" *Amer. Midland Naturalist*, Vol. XV (1934), pp. 752–59.

[14] *Op. cit.*

[15] "Post-hurricane Sea Shore Observations," *op. cit.*, Vol. XVIII (1937), pp. 426–34.

invertebrate shells embedded in them. From these and other examples the inescapable conclusion is reached that in the genesis of lake balls or sea balls there commonly is diversity of organic inception but always similarity of physical conditions requisite for their later stages of development.

Since these conditions are rarely satisfied today, sea balls are uncommon sedimentological phenomena. Little wonder, therefore,

FIG. 6.—Heterogeneous meshwork of siliceous fossils characteristic of "Niagaran sea balls" after their partial reduction with acid.

that they have not been discovered previously in marine sediments, although Kindle has seen in "some 'coal balls' " structures analogous to "some modern 'lake balls' " of plant origin. The Niagaran sea balls undoubtedly owed their inception to the subfibrous character of the bryozoans (Fig. 6), which dominate the fauna found in them. The physical conditions responsible for their growth into sea balls have been hypothecated previously.

At the present time good specimens of the nodules or sea balls are known only from a sporadic occurrence in a restricted zone in a single large quarry (Federal Stone Company). Later investigations may reveal a much wider distribution but, judging from the re-

stricted modern geographic range of sea balls, this is only a possibility, not a probability.

MODE OF TREATMENT OF NODULES

The siliceous nature of the fossils in the sea balls permits solution with acid to dissolve the dominantly calcareous matrix. In the laboratory the rock surrounding a ball is removed as carefully as possible and the cleaned specimen is then placed on a pedestal in a beaker and subjected to a dilute hydrochloric acid bath. It has been determined experimentally that a 5 per cent acid solution produces the best results since the vigor of reaction is thus reduced to a minimum and does not in itself noticeably injure the more delicate fossils. About 10 days are required to dissolve an average ball with acid of this concentration. The pedestal serves to keep the undissolved portion of the ball from crushing the more fragile forms that are set free and fall to the bottom of the beaker. Some score of balls have been treated to date. The intimate mixture and random orientation of the fossils in a partly reduced nodule are illustrated in Figure 6.

After the solution of a nodule is completed, the material is washed and decanted and the residue is dried and carefully sieved into four grade sizes for convenience of examination. The sizes selected are the coarse fraction that is retained on a standard twenty-mesh screen, the fractions retained on the forty and sixty meshes, and all material that passes the sixty-mesh screen.

Several of the nodules have been quantitatively studied to determine if a definite relationship exists between the fossil content and the mass of different specimens. The nodules were weighed before acid reduction and the weight percentages of the residues were computed. Three small nodules that weighed 58, 190, and 227 gm. contained respectively 4.5, 9.2, and 12.0 gm. of fossil material. The equivalent weight percentages are 7.8, 4.8, and 5.3. A larger nodule that weighed 628 gm. contained 113 gm., or 18 per cent fossils by weight. From a comparison of these figures it is evident that there is no very definite ratio between the weight of the nodules and the fossils. Argillaceous mud and silt easily recovered account for at least 3 per cent by weight of the above specimens. In addition,

much of the $CaCO_3$ reduced was in the form of a lime mud. In a sense, then, the nodule matrix and laminae have something of the sedimentological characteristics of a water lime.

COMPOSITION AND GENERAL ASPECTS OF FAUNA

The original hard parts of the fossils have been so well replaced by silica that most of the detail of shell structure is preserved. The fossils are mostly small, their long dimensions ranging from less than 1 to more than 20 mm. As an approximation the average specimen is probably less than 8 mm. in greatest dimension.

Most of the more common Paleozoic classes of invertebrates occur in the fauna of a single nodule, but bryozoans, brachiopods, and corals are by far the most abundant of the groups represented. Gastropods, crinoids, and ostracodes follow more or less in that order of diminishing abundance. In addition, the following groups have some representation: worms, pelecypods, blastoids, sponges, holothurians or echinoids, cephalopods, trilobites, and some foraminifera. Disarticulated echinoderm parts are common in every sample.

The following is a brief synopsis of the fauna:

Protozoa.—A few possible lagenoid types of foraminifera and abundant rodlike, probable *Astrorhizidae* occur; *Radiolaria* may be represented by some more or less indeterminate spicules.

Porifera.—No complete sponges have been found, but spicules are common in the smaller-size fractions. Two types are recognized: a hexactinellid spicule of which four rays are in one plane and two in another, the planes being at right angles to one another, and an *Astraeospongia* type, which has six rays symmetrically spaced in a single plane with two shorter rays at right angles.

Coelenterata.—The corals are represented by species of both individual and colonial types. A small, delicate tetracoral displays all stages of growth and presents a well-defined sequence of septal development. Some colonial tabulate species belong to the genus *Favosites*, and some *Auloporidae* also are represented.

Vermes.—A few worms, which belong to the genus *Spirobis*, and some types of uncertain affinities, are present.

Echinoderma.—Several articulated basal plates, apparently of the

common Silurian blastoid *Troostocrinus reinwardti*, have been observed. Crinoids are chiefly represented by numerous plates and ossicles, but at least three species are recognized from complete dorsal cups. The maximum dimensions of entire individuals do not exceed 2 mm.; nevertheless, the forms are apparently mature. Frondlike plates that probably belong either to the echinoid or the holothurioid group are common.

Bryozoa.—Bryozoans are the most abundant fossils in the fauna. Clublike individuals, ramose, encrusting, and fenestelloid types are common in every nodule examined. Detailed investigation is not complete, but species have been assigned with certainty to the genera *Ceramoporella* and *Rhinidictya;* the fenestelloid types are new species.

Brachiopoda.—The brachiopods are second only to the bryozoans in relative abundance. Several common Niagaran species are present, notably *Parmorthis waldronensis* (*Dalmanella elegantula*), *Spirifer radiatus*, *Atrypa nodostriata*, *Rhipidomella hybrida*, and *Rhynchotreta cuneata* var. *americana*. In addition, there are some new genera and species.

Mollusca.—The genus *Conocardium* is the only pelecypod represented, but several genera of gastropods, especially *Diaphorostoma* and *Hormotoma*, occur. A peculiar new patellid type with a strongly ribbed eccentric cone is common also. Cephalopods are restricted to a few poor orthoconic fragments.

Arthropoda.—Fragments of trilobites, notably free cheeks and spines, have been observed, but ostracodes as individuals constitute a much more important part of the fauna. They seem, however, to be confined to two species. A new species of *Beyrichia* is abundant, with both male and female individuals well represented.

PECULIARITIES OF THE FAUNA

It may be inferred from some of the foregoing statements that the fauna is either dwarfed or composed mainly of young individuals. Some salient features seem to indicate that it is dwarfed. For example, brachiopod species which ordinarily attain dimensions of as much as 30 mm. are represented by individuals that do not exceed 10 mm., yet they are ventricose, have a mature pedicle arrangement,

are marked with several seasonal growth lines, and have features of shape proportional to the larger and more commonly known forms. Crinoids, in which the calyces do not measure more than 2 mm., have complete cyclic arrangements of basal and radial plates with well-developed arm sockets, characteristic of mature representatives of related species. Cup corals display as many as 40 septa, yet the length of the corallum of the largest individual does not exceed 12 mm. The pelecypod *Conocardium* is represented by several individuals, of which none has a length of more than 8 mm., yet their number of striations is comparable to that for larger forms in other faunas. None of the many gastropods in the fauna is as long as 10 mm., yet most have the same number of whorls as their more characteristic relatives in other sediments.

Unfortunately for any very simple dwarfing concept, the bryozoans do not appear to be markedly reduced in size, although they are somewhat smaller than the average of their relatives in other faunas. The ostracodes, as characterized by the genus *Beyrichia*, also are relatively large, and commonly have a length of more than 3 mm. This is normal, or larger than normal, for the genus in other deposits. Thus, if dwarfing has occurred, the environmental agencies which produced the dwarfed condition were either not operative or not uniformly effective for all the invertebrate groups.

If the storm theory is accepted, then the amount of sedimentation concurrent with the formation of the nodules should be some indication of the duration of the storms. Since the nodules are known at present to have a stratigraphic range of as much as 3 feet, this amount of deposition in an extensive sea with adjacent lands of low relief should, according to our present concepts, represent a rather extended period of time. The introduction of mud and silt into a normally clear sea, if extended over several seasons, could conceivably disturb the ecological equilibrium of the marine organisms and result in a dwarfed fauna. De Lapparent,[16] for instance, points out that the sediment from the Nile apparently has exercised such a dwarfing effect on the organisms of the eastern Mediterranean Sea. Free-swimming forms such as the ostracodes, however, may

[16] A. de Lapparent, *Tráite de géologie* (5th ed.; Paris: Masson & Cie, 1906), pp. 132–33.

well be invulnerable to sedimentological factors operative largely on the sea floor. In fact, ostracodes do not readily respond with a size diminution to conditions which cause dwarfing in other classes. This point is well shown by the Tully representatives of the group which, as Loomis[17] has demonstrated, are markedly unaffected members of the Tully dwarfed fauna. The approach to size normalcy in a number of the bryozoans is more difficult to explain, but it is possible that some of them lived and died before the dwarfing agencies became effective, since a number of the specimens show evidence of having been abraded. Semper,[18] however, states that many typically marine bryozoans also can survive in fresh water. This fact seems to indicate that bryozoans have considerable resistance to adverse environmental conditions. Possibly, therefore, the bryozoans of the sea balls were not so readily affected by the dwarfing agencies as the other forms.

Despite the foregoing partial exceptions it is clear that any dwarfing process should naturally affect, at least to some extent, all forms living in the environment. Hence, a dwarfed fauna should occur in the beds enclosing the nodules as well as in the sea balls themselves. Unfortunately, the process of dolomitization has obliterated or rendered undeterminable any fossils in the dolomitic strata adjacent to the balls.

A more detailed description of the nodules and their fauna will be released for publication shortly by the junior author.

ACKNOWLEDGMENTS.—We are indebted to Professor J Harlen Bretz and Dr. E. H. Stevens for calling this problem to our attention and to Dr. Homer J. Smith of Tulsa University for much of the early work in reducing the nodules and sorting the residues. We are also obligated to the late A. W. Slocom, who introduced the senior author to the subject of "lake balls."

[17] F. B. Loomis, "Dwarf Fauna of the Pyrite Layer of the Tully Limestone," *N.Y. State Mus. Bull. 69* [(1902], 1903), p. 899.

[18] Karl G. Semper, *Animal Life as Affected by the Natural Conditions of Existence* (New York: D. Appleton & Co., 1881), p. 148.

AN EXHUMED EROSION SURFACE IN THE JEMEZ MOUNTAINS, NEW MEXICO

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ABSTRACT

The plateau-like surface of San Pedro Mountain in the northwestern part of the Jemez Mountains of New Mexico is a remnant of an exhumed erosion surface. It is overlain in places by a thin veneer of chert, here named the Pedernal chert, with underlying gravel. The chert and gravel are a part of a thin wedge of Tertiary beds which thickens eastward and extends across the northern spurs of the Jemez Mountains. The rocks of the wedge are correlative with the El Rito formation, Abiquiu tuff, and Santa Fe formation, found in the Rio Grande depression east of the border faults of that trough. The stratigraphy and structure of the Tertiary sedimentary rocks of the wedge reveal in some detail the history of the deformation of the Rio Grande depression. The mountains formed during the Laramide revolution were reduced in early Tertiary time to a low-lying erosion surface or peneplain. A slowly subsiding basin then formed and was maintained until late Tertiary time, when it was broken by the normal faults which now bound the Rio Grande depression, leaving some of the Tertiary basin deposits on the western mountain block. Erosion of the larger part of these Tertiary rocks has left exposed on the summit of San Pedro Mountain a large area of the exhumed early Tertiary peneplain.

INTRODUCTION

The area to be considered in this paper lies in the northern portion of the Jemez Mountains of New Mexico. This range makes up the southern extremity of the western prong of the Southern Rocky Mountains. On its north side it is connected to the main group of the Southern Rockies by a narrow belt of plateaus of lower elevation underlain by gently folded rocks which extend northward into Colorado. The Jemez Mountains overlook the sage-covered Colorado Plateau on the west and the low, fertile Rio Grande Valley on the south and east. Thus on all sides they stand up as a prominent and large highland mass.

The geologic mapping of the Jemez Mountains has as yet proceeded only to the stage of reconnaissance work. The general geology

has been briefly described by Darton.[1] The western edge of the mountains where they border the plateau has been described in more detail by Renick.[2] Though little physiographic work has been done in the mountains, the Tertiary history of the Rio Grande Valley and the plateau to the west of the mountains is fairly well known. Bryan, Bryan and McCann, Smith, and Cabot have described in some detail the deposits of the Rio Grande Valley.[3] They have shown that this valley is structurally a great graben or depression separating the two prongs of the Southern Rockies or, in other words, that the Rio Grande depression is a graben of the type found in the basin and range province to the south which extends northward into the Southern Rockies, splitting them into two prongs. The downfaulting which formed the Rio Grande depression occurred in Pliocene and Pleistocene time, accompanied by extensive alluvial deposition. But earlier, in Tertiary time, initial basin deposits

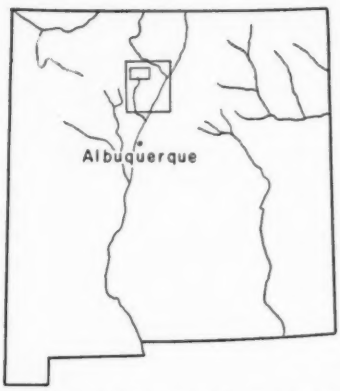

FIG. 1.—Index map of New Mexico, showing location of area studied in the field (inner square) and area shown in maps of Figures 2 and 10 (outer square).

[1] N. H. Darton, "Red Beds and Associated Formations in New Mexico," *U.S. Geol. Surv. Bull. 794* (1928).

[2] B. C. Renick, "Geology and Groundwater Resources of Western Sandoval County, New Mexico," *U.S. Geol. Surv. Water Supply Paper 620* (1931).

[3] Kirk Bryan, "Geology and Ground Water Conditions of the Rio Grande Depression in Colorado and New Mexico," *Regional Planning*, Part VI; *The Rio Grande Joint Investigation in the Upper Rio Grande Basin in Colorado, New Mexico, and Texas, 1936–1937* (Washington: U.S. National Resources Committee, 1938), Vol. I, Part II, sec. 1, pp. 197–225, Figs. 45–52 incl.; Kirk Bryan and F. T. McCann, "The Ceja del Rio Puerco: A Border Feature of the Basin and Range Province in New Mexico," *Jour. Geol.*, Vol. XLV (1937), pp. 801–28, and Vol. XLVI (1938), pp. 1–16; "Successive Pediments and Terraces of the Upper Rio Puerco in New Mexico," *ibid.*, Vol. XLIV (1936), pp. 145–72; H. T. U. Smith, "Tertiary Geology of the Abiquiu Quadrangle, New Mexico," *Jour. Geol.*, Vol. XLVI (1938), pp. 933–65; E. C. Cabot, "Fault Border of the Sangre de Cristo Mountains North of Santa Fe, New Mexico," *Jour. Geol.*, Vol. XLVI (1938), pp. 88–105.

were laid down in the region before any known, or as yet proved, faulting began. As shown in this study, the sedimentation in these initial basins was accompanied by gentle downwarping of the area now occupied by the Rio Grande Valley. The writers have been able to trace the initial basin deposits of pre-Pliocene age from the Rio Grande depression westward into the northern border of the Jemez Mountains and have found them as far as 20 miles west of the present depression on the high plateau of San Pedro Mountain, at the northwest corner of the Jemez Mountains and adjacent to the Colorado Plateau. They are unfaulted but are gently deformed and reveal in some detail the history of the basin in which they were deposited.

Underneath the initial basin deposits is an erosion surface of low relief which has become exhumed on San Pedro Mountain. A great sheet of chert, here named the Pedernal chert, with thin underlying gravel, still remaining on portions of the erosion surface, has held up the mountain surface by its great resistance to erosion. Its relatively recent dissection accounts for the gently undulating topography of the broad and unusual mountain top. At the time when the present study was planned it was thought that the surface of this mountain might be a remnant of the extensive San Juan peneplain of the San Juan Mountains.[4] It is, however, an exhumed peneplain older than the San Juan peneplain, and remnants of the latter may not be preserved in the Jemez Mountains.

Thus two separate problems are involved: (1) the nature of the Tertiary deformation of the Rio Grande depression and the uplift of the Jemez Mountain mass and (2) the existence and nature of a remnant of a once-extensive Tertiary peneplain. The two problems are closely related, for the peneplain is dated by the remnants of the Tertiary beds that overlie it, and the structural history of the region can be understood from an analysis of the deformation of these beds and of the peneplain itself.

GENERAL DESCRIPTION OF THE JEMEZ MOUNTAINS

Although no detailed work has been done on the geology of the interior of the Jemez Mountains, the writers have prepared a sketch

[4] W. W. Atwood and K. F. Mather, "Physiography and Quaternary Geology of the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 166* (1932).

map of the whole region, compiled from published and unpublished reports of several workers (Fig. 2).

In the southern portion of the Jemez Mountains, close to the Rio Jemez, the downfaulted late Tertiary rocks of the Rio Grande depression lie adjacent to the Mesozoic rocks of the Colorado Plateau. They are separated by a northward-trending normal fault. About 10 miles north of the southern edge of the map the line of en echelon faults swings to the east (see also Fig. 10). At this point the Mesozoic rocks of the plateau on the west become separated from the late Tertiary rocks of the Rio Grande depression by a complex group of sedimentary and igneous rocks ranging in age from the pre-Cambrian to the Quaternary. These make up the Jemez Mountains. Along the northern boundary of the Jemez Mountains the line of en echelon faults bounding the Rio Grande depression again swings to the west at the town of Abiquiu, forming what is known as the Abiquiu re-entrant.

The Jemez Mountains themselves are a complex group. They are bounded on the west side by a north-south high-angle thrust fault of Eocene age, known as the Nacimiento thrust. Along this fault the pre-Cambrian, Paleozoic, and Mesozoic rocks of the Sierra Nacimiento are thrust over the Mesozoic rocks of the Colorado Plateau. The Sierra Nacimiento is a long but narrow group of peaks, forming a serrate ridge, rising to elevations of 9,000 feet at the southern end and culminating in the broad plateau of San Pedro Mountain at the northern end, which rises to an elevation of over 10,600 feet. East of the Sierra Nacimiento, Paleozoic rocks rest unconformably with slight dips to the east on the pre-Cambrian core of the ridge. Successively younger sedimentary rocks are exposed to the east but they become buried by the great blanket of volcanic rocks of Tertiary and Quaternary age which surrounds Cerro Redondo, an ancient and lofty volcanic cone.

The Rio Grande depression is largely underlain by the Santa Fe formation, which includes alluvial sand and gravel with interbedded basaltic lava. In places, and especially in the Abiquiu re-entrant, there are older basin deposits of Tertiary age. In this re-entrant the Santa Fe formation is underlain by the thick Abiquiu tuff, which is in turn underlain by the El Rito formation.[5] Now the

[5] Smith, *op. cit.*

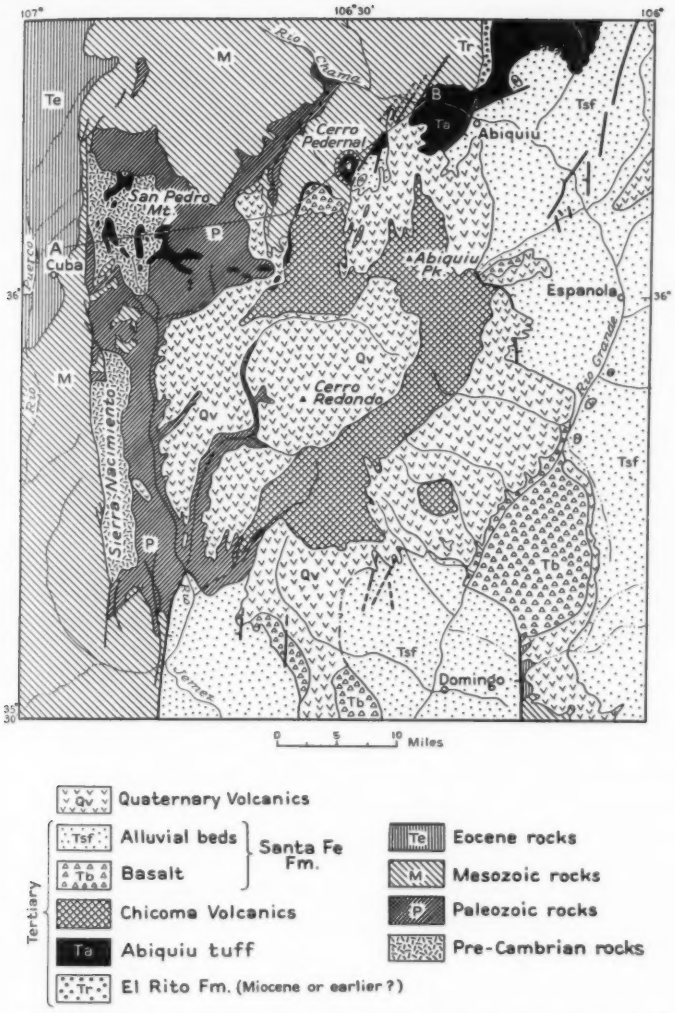

FIG. 2.—Generalized geologic map of the Jemez Mountains. The normal faults of the Rio Grande depression are shown by heavy black lines, whereas the faults of the Nacimiento are shown by lighter black lines. Compiled from published maps by Darton, Bryan and McCann, and Smith. The Domingo area is based on an unpublished map loaned by Professor Kirk Bryan and Dr. J. E. Upson. The volcanics of the northern portion of the Jemez Mountains are based in part on an unpublished map loaned by Professor E. S. Larsen, Jr. The Tertiary sedimentary rocks between Cerro Pedernal and San Pedro Mountain are based on the writers' field work, described in this paper.

bounding faults of the Rio Grande depression usually define the limits of the basin deposits, which are absent in the uplifted highland blocks. However, in this area, although the faults pass just east of Cerro Pedernal, the Tertiary basin formations may be traced westward in the northern spurs of the Jemez Mountains to San Pedro Mountain. They lie unconformably on the pre-Cambrian, Paleozoic, and Mesozoic rocks, and underlie the Quaternary volcanics of the mountains, as shown in the structure section (Fig. 3).

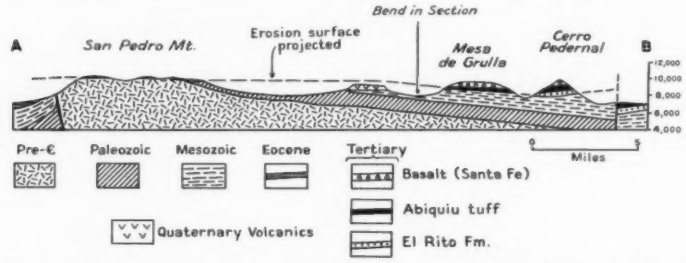

FIG. 3.—Section across the northern part of the Jemez Mountains. Vertical exaggeration is about three times. The line of the section is indicated in Figure 2 by the line *A–B*.

TERTIARY SEDIMENTARY ROCKS OF THE JEMEZ MOUNTAIN BLOCK

The writers first encountered these Tertiary basin deposits on San Pedro Mountain (see Fig. 4) where in many places a thin veneer of chert and underlying gravel overlies the broad plateau-like surface that forms the summit of this mountain. The chert, because of its remarkable purity, distinctive appearance, and widespread distribution, is here named the Pedernal chert after Cerro Pedernal, a mountain whose Spanish name means "flint." It presumably was so named from the outcrops of chert long quarried by the Indians. The chert was mapped on San Pedro Mountain and its horizon was followed eastward into the high mesas of the northern part of the Jemez Mountains (see Fig. 4). It passes laterally without losing continuity into poorly consolidated conglomerate, sandstone, shale, and tuff, which in Cerro Pedernal are obviously part of the Abiquiu tuff. The Pedernal chert is thus only a member of this formation.

Cerro Pedernal is included in the Abiquiu quadrangle, and its geology was mapped by Smith,[6] who showed that the Tertiary rocks

[6] *Ibid.*

which hold up its summit and upper slopes are correlative with the Tertiary basin deposits of the Abiquiu re-entrant. Therefore the Pedernal chert and the other rocks of the wedge of sedimentary material extending west from the Cerro can be divided into the formations recognized in the Abiquiu quadrangle and given an approximate geologic date.

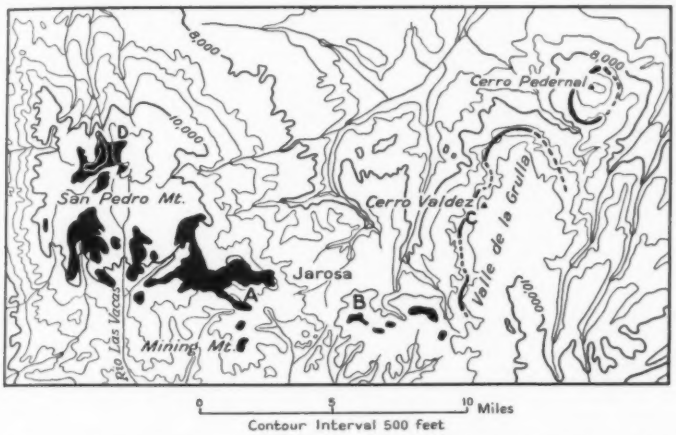

FIG. 4.—Contour map of the northern portion of the Jemez Mountains, reduced from a map of the U.S. Forest Service. The areas of Pedernal chert are shown in black.

The El Rito formation of reddish shale, arkose, and tuff in Cerro Pedernal lies unconformably on the Cretaceous Dakota sandstone. Its age is late Eocene, Oligocene, or early Miocene, but more probably late Eocene or Oligocene.[7] It is not found west of Cerro Pedernal and probably pinches out under the Valle de la Grulla.

The Abiquiu tuff in this region may be divided into three members: a basal gravel member, the relatively thin Pedernal chert member, composed of white chert and white siliceous limestone, and an upper or tuff member, which farther east in the depression makes up most of the formation.

The basal gravel member consists of coarse gravel interbedded with coarse sand on Cerro Pedernal; it becomes relatively fine grained to the west, containing interbedded shale on the east side of

[7] *Ibid.*, p. 940.

San Pedro Mountain. It is about 220 feet thick on Cerro Pedernal and thins westward to 60 feet on the east side of San Pedro Mountain. It is missing on the west side of this mountain. On Cerro Pedernal the basal gravel rests on the El Rito formation, but west of there it forms the base of the Tertiary basin deposits. On San Pedro Mountain it rests on the pre-Cambrian or Paleozoic rocks.

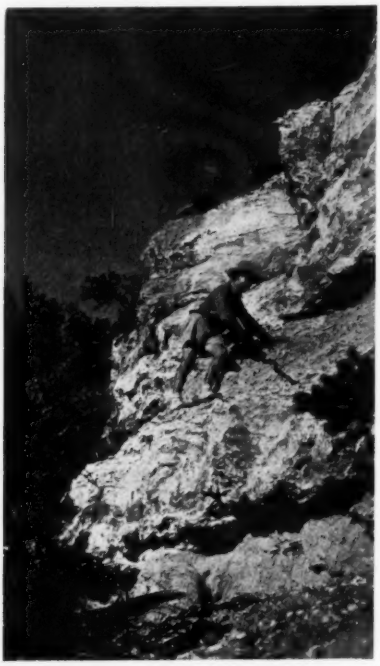

FIG. 5.—Outcrop of Pedernal chert on the north side of Pedernal Peak

The Pedernal chert member of the Abiquiu tuff is a good horizon marker, extending across the area from Cerro Pedernal to the west side of San Pedro Mountain. The chert is almost everywhere associated with a bed of siliceous limestone which usually becomes more siliceous from bottom to top, grading into chert. In places there is chert at the bottom of the limestone bed. This member is about 25 feet thick on Cerro Pedernal (see Fig. 5). On Cerro Valdez two thin

beds of chert are separated by about 100 feet of soft material covered with landslide debris. At locality *B* (Fig. 4) south of Jarosa two beds of chert are separated by about 40 feet of limestone. The member consists of one bed of chert and limestone on the east side of San Pedro Mountain. On the west side of this mountain only chert was found, although good exposures are rare in this region. Thus the Pedernal chert member appears to thin somewhat on San Pedro Mountain. It may once have extended farther to the west. The chert is remarkably pure and is distinguishable by its color, which

FIG. 6.—Indian quarry pits in Pedernal chert on north side of San Pedro Mountain

is usually a pearly white though flecks and stains of red, yellow, and black, apparently the result of weathering, are common. It is massive and made excellent material for artifacts[8] (see Fig. 6). In most places it is easily mapped, even on the flat surface of San Pedro Mountain, for it weathers into rounded blocks which project up through the soil.

The upper or tuff member of the Abiquiu tuff is found only on Cerro Pedernal, where it is probably about 400 or 500 feet thick.[9] It

[8] Artifacts and quarries associated with the chert are described by Kirk Bryan, "Prehistoric Quarries and Implements of Pre-Amerindian Aspect in New Mexico," *Science*, Vol. LXXXVII (1938), pp. 343–46.

[9] Smith, *op. cit.*, p. 948.

may be present also on Cerro Valdez, but if so it is covered by landslide debris.

Thus the Abiquiu tuff thins westward and becomes finer westward. The Pedernal chert member contains the finest material and extends farthest to the west. The tuff member has probably been removed by erosion west of Cerro Valdez. The age of the Abiquiu tuff is not known with certainty, but it is thought by Smith to be Miocene.[10]

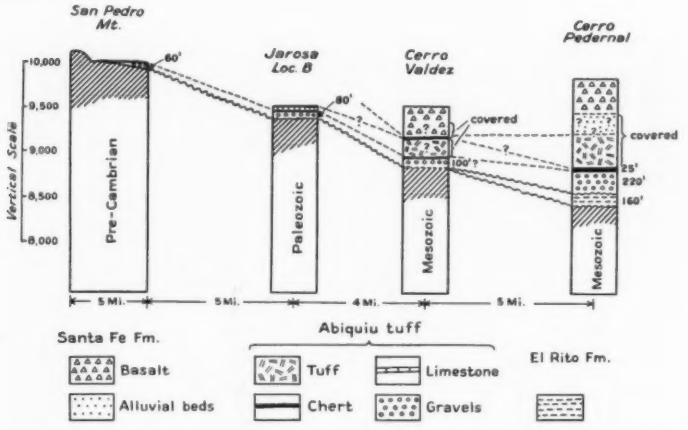

FIG. 7.—Columnar sections showing the stratigraphic relationships of the Tertiary rocks in the northern portion of the Jemez Mountains.

The Santa Fe formation of late Miocene and Pliocene age[11] overlies the upper or tuff member of the Abiquiu tuff on Cerro Pedernal. It is composed of interbedded basalt and alluvial sand, which is obscured by talus. Basalt flows of the Santa Fe formation cap Cerro Pedernal and extend westward as far as Cerro Valdez. The formation is not found west of that point.

Thus the Tertiary rocks of the region occur in a wedge, thinning westward from 1,400 feet on Cerro Pedernal to nothing at the summit of San Pedro Mountain, as represented diagrammatically in Figure 7. The El Rito formation of Eocene to Miocene age pinches

[10] *Ibid.*, p. 958. [11] *Ibid.*, p. 957.

out rapidly, probably under the Valle de la Grulla. The lower part of the Abiquiu tuff of Miocene age, consisting mostly of gravel, thins, becomes finer, and pinches out on San Pedro Mountain. The higher portions of the Abiquiu tuff above the chert and the Santa Fe formation do not extend beyond Cerro Valdez. Both formations may and probably did originally thin to the westward, but considerable thicknesses must have been eroded from the western part of the area here considered.

FIG. 8.—View from the summit of San Pedro Mountain, looking westward

THE EXHUMED EROSION SURFACE ON SAN PEDRO MOUNTAIN

San Pedro Mountain is the culmination of the north-south Nacimiento Ridge but is wholly different in appearance from the surrounding mountain areas. It has the appearance of being the remnant of an ancient peneplain. In fact, a traveler to the summit can imagine he is on the Canadian Shield and not in New Mexico (see Fig. 8). The mountain surface as well as the other areas in which the Pedernal chert is found is shown in Figure 4.

The mountain rises steeply from the west side; from the north its rounded flanks present a more gradual ascent to what appears to be a high plain. This plain drains principally southward into the tributaries of the Rio Las Vacas, whose waters are carried to the Rio Grande by the Rio Jemez. Fingers of the plain extend to the southeast forming flat, poorly defined subdivides and are rimmed by grayish cliffs of moderate height. The summit of San Pedro Mountain is slightly over 10,600 feet in elevation; there are over 4 or 5 square

miles within 150 feet of this height and some 40 square miles above 10,000 feet.

To the north, east, and west the drainage of this mountain may be described as abrupt; the streams fall away with steep gradients, usually in steep-walled canyons and gorges. In contrast, the southerly drainage to the Rio Las Vacas begins on the upland plain in grassy flats, many marshy, some over $\frac{1}{4}$ mile wide and five times as long. These meadows, fringed with uniform and regular stands of spruce,

FIG. 9.—View of Pedernal Peak from the north side of the Chama Valley

are known as the San Pedro Parks and are much admired for their natural beauty. Through these meadows the headstreams meander sluggishly. Between the park areas, as between the streams somewhat lower down, the divides are not at all sharp but preserve the gently sloping surface. There are places where the heads of streams flowing north and west have broken through the low subdivides and have captured the waters formerly carried by the Vacas, so that their waters fall off the mountain to the west and north in steep little canyons which feed the Rio Puerco and the Rio Gallina. Though the direction of drainage is to the south and west, the slope of the mountain is in general to the southeast. Between the parklike valleys the broad flat divides covered by thick spruce forests preserve this slope. Most of the area is underlain by pre-Cambrian granite and granite gneiss, but preserved on portions of the divides are remnants of the

great sheet of Pedernal chert which once covered the whole mountain except where a few monadnocks stood above the general level. The valleys have cut a short distance into the chert and the erosion surface underneath it, giving the topography of the summit area, above the precipitous slopes, the appearance of a gently rolling plain (Fig. 8).

The highest points of the summit area are low granite knobs which rise above the level of the chert and are thus monadnocks above the erosion surface on which the chert was deposited.

The Pedernal chert and gravel, both members of the Abiquiu tuff, have been shown to overlie unconformably the pre-Cambrian rocks and Paleozoic rocks which dip outward in all directions from the summit of the mountain. Therefore this unconformity must mark an erosion surface, and it is a surface of very low relief. The exhumed portion of the erosion surface is exposed over only a small area on San Pedro Mountain, where the chert has been stripped off and has left the pre-Cambrian granite and granite gneiss exposed. The chert and its underlying gravel are, however, only 20–80 feet thick on San Pedro Mountain so that it is indeed the topography of the ancient erosion surface which is now revealed for many miles on San Pedro Mountain.

The erosion surface bevels the Paleozoic and Mesozoic rock strata of the Jemez Mountains, which were folded and faulted in Eocene or post-Eocene time.[12] Therefore the surface must be later than early Eocene. As the Abiquiu tuff rests on the erosion surface it must be pre-Abiquiu (Miocene) in age. But, as can be seen in the columnar sections of Figure 7, the erosion surface on San Pedro Mountain might be projected either into the unconformity between the Abiquiu tuff and the El Rito formation or into the unconformity between the El Rito formation and the pre-Tertiary rocks. This possible choice in time can probably be best interpreted to indicate that erosion was continuous in the region of San Pedro Mountain during the deposition of the El Rito formation in the region of Cerro Pedernal and the Abiquiu re-entrant. The western margin of the basin of sedimentation was presumably low-lying and virtually in a state of peneplanation during the period of sedimentation, or in other words

[12] Renick, *op. cit.*, pp. 54 and 55.

for most of Oligocene and Miocene time. However, there must have been hilly areas and streams with relatively steep gradients, else the coarse materials of the El Rito and basal Abiquiu formations could not have reached the basin of sedimentation. Just where these areas were located is not now known, but it appears that the region immediately west of Cerro Pedernal including San Pedro Mountain was not one of them.

DEFORMATION OF THE EROSION SURFACE

The Pedernal chert in most places where examined is a replacement of, or is associated with, limestone which is in places over 40 feet thick. This limestone was presumably deposited on a surface which was flat or nearly so and with a very low gradient. But at the present time the chert and limestone have a slope in some places of over 200 feet per mile. Therefore the Pedernal chert has been deformed and the erosion surface beneath it has also been deformed. On Cerro Pedernal the chert is over 1,000 feet lower than it is on San Pedro Mountain. It has been shown in the preceding paragraph that while the El Rito formation was being deposited at Cerro Pedernal, the San Pedro Mountain region was probably undergoing erosion. This suggests that some of the warping may have occurred before the Pedernal chert was deposited. In other words an initial basin formed in the region of Cerro Pedernal (and eastward in the Abiquiu reentrant where the El Rito formation is still thicker). This basin was broad and shallow. Probably alternate deposition and erosion went on during a long interval in Tertiary time without much change in level. But in Miocene time the basin became accentuated and greater thicknesses of alluvial sediments were deposited in the region. Eventually the basin was greatly downwarped and broken by normal faults.

The Pedernal chert, whose slope now measures at least the minimum deformation since Miocene time, has been contoured (see Fig. 10). The contours run in a northeasterly direction parallel to the trend of the faults which bound the Rio Grande depression. Although the area that can be contoured is small, this parallel relationship is clearly shown. The original or initial basin of deposition, which developed by the warping of the early Tertiary erosion sur-

face, was wider than the present Rio Grande depression, its western border being near or west of San Pedro Mountain. It was slightly downwarped during El Rito time (early Tertiary) and was probably

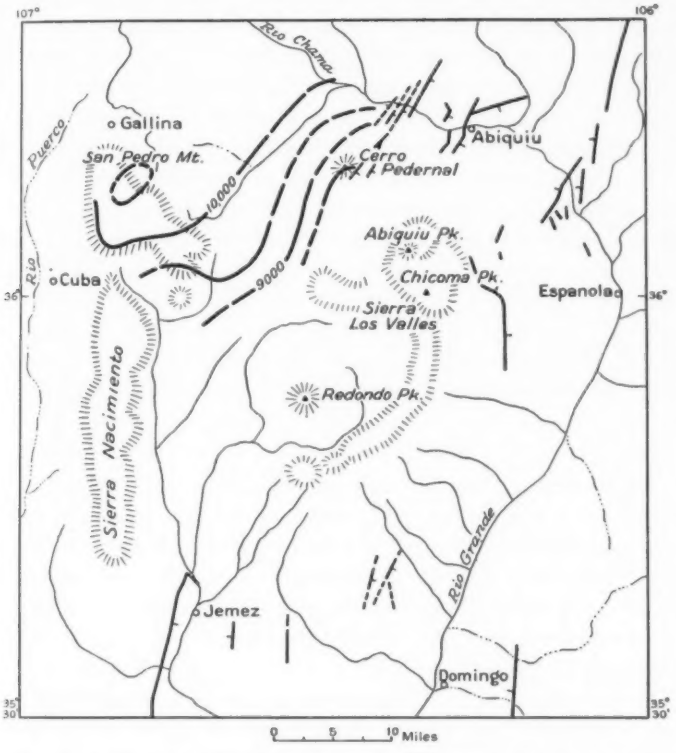

FIG. 10.—Map of the Jemez Mountains showing the normal faults of the western portion of the Rio Grande depression. The contour lines are drawn on the Pedernal chert and show approximately the amount of deformation of the western mountain block of the Rio Grande depression.

accentuated during Abiquiu time (Miocene), but most of the warping occurred after the formation of the Pedernal chert (Abiquiu), and some of it may have been contemporaneous with the Pliocene faulting which marked out the present Rio Grande depression. This

faulting broke the original basin of deposition along a line far to the east of its western border, leaving the western portion of the basin high on the mountain block west of the Rio Grande depression.

It is possible, however, to speculate as to how much of the original Tertiary basin of sedimentation was cut off and uplifted as part of the rising mountain block. Bryan has traced the course of the Pliocene Rio Grande from a point south of Belen as far north as the Española region. He has evidence that north of Española it flowed from the northwest and its course probably was near Cerro Pedernal.[13] If this picture is correct, probably the whole western half of the original basin of sedimentation was cut off from the present Rio Grande graben by the late Pliocene normal faults. The sediments of this western portion of the basin, which may have had a considerable thickness, must have rested on the mountain block at the close of faulting. This unconsolidated material has been nearly all removed and the present summits of the western mountain block rise to, or have been eroded below, the floor of the Tertiary basin of sedimentation.

CONCLUSION

The parallel relationship between the axes of deformation of the Abiquiu tuff and the marginal faults of the Rio Grande depression indicates that the two phenomena represent different manifestations of the same process—the sinking of a great graben relative to its bounding highland blocks. The stratigraphy of the lenticular basin of sedimentary rocks, which still persists now on the Jemez Mountain block, indicates that at least some of the warping preceded the faulting. That this wedge of sediments must represent only a portion of a much wider basin whose center in Miocene time may have been east of Cerro Pedernal is shown not alone by the fact that the Abiquiu tuff thickens eastward, but also by the fact that its basal sediments become coarser to the east. In Santa Fe (Pliocene) time, the center of the basin was probably just west of Cerro Pedernal, and the post-Santa Fe normal faulting lifted the whole western portion of the basin. Most of the basin deposits have since been eroded from the mountain block. The early history of the eastern

[13] Bryan, "Geology and Ground Water Resources . . . . ," *op. cit.*, pp. 205–7.

portion of the basin is not known and there is no reason to suppose that it was precisely similar to that of the western portion.

The exhumed erosion surface on San Pedro Mountain reveals the existence of a long period of erosion in early Tertiary time, during which the mountains formed during the Eocene folding and thrust-faulting were beveled to a surface that was in places a remarkably smooth plain. It was after the development of this erosion surface that the basin of the Rio Grande became established, to be first gently downwarped and gradually accentuated during the remainder of Tertiary time.

ACKNOWLEDGMENTS.—Much of the territory studied would have been inaccessible without the pack animals and camp equipment furnished by Los Alamos Ranch School. The mapping was simplified by the use of forest reconnaissance maps furnished through the co-operation of the United States Forest Service. Thanks are due to F. C. Hack for his interest and support, to Mrs. F. S. Church, Theodore, Allen, and Hugh Church, and Steve Edsall for assistance in camp and in the field. The writers are grateful to Professor M. P. Billings and E. C. Cabot for co-operation and suggestions during the preparation of the material; but the general success of the expedition stems from the imagination and guidance of Professor Kirk Bryan, both in the field work and in the preparation of the manuscript.

STRUCTURAL FEATURES OF A LANDSLIDE NEAR GILROY, CALIFORNIA

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ABSTRACT

Structures produced by plastic flow in a recent large landslide near Gilroy, California, are strikingly similar to certain structures produced on a larger scale in the deformation of rocks. Fracture patterns along strike-slip faults, and successive normal faults due to tension and rotation, are especially well displayed. Well-defined grabens were formed where the slide overrode small hills. Local compression has produced fractured anticlinal ridges. Relations of these structures to the slide's movement and to the original topography are discussed with the aid of detailed maps.

Recent movement on a landslide in the hills south of Gilroy, California (on the north slope of the Lomerias Muertas, lat. 36°54′ N. and long. 121°32′ W., San Juan Bautista quadrangle), has developed remarkably well certain structures characteristic of plastic flow—structures resembling on the one hand those encountered in large-scale deformation of rocks, on the other hand those produced in laboratory clay-box experiments. The following pages present an analysis of these structures in terms of the forces and movements responsible for them.

The Lomerias Muertas are a group of low hills in the California coast ranges, rising by gentle slopes some 800 feet above the Pajaro Valley. They are grass-covered grazing land, barren of trees except along the major stream courses. The Gilroy slide is an unusually large example of a phenomenon which is very common in such localities during wet weather: a movement of soil, mantle rock, and soft bedrock rendered plastic by saturation with water. Scars on adjacent slopes testify to similar landslides in other years, their number suggesting that much of the material of the present slide had taken part in previous movements.

According to local cattlemen, large cracks near the top of the slope were first noticed on March 5, after a few days of moderately heavy rain. The principal movement occurred in the early morning hours of March 9 and was sufficiently rapid and noisy to cause a near-panic

in the surrounding countryside. During several days the motion subsided, practically ceasing by March 13.

Widespread and still persistent rumors that gas explosions started the slide were carefully checked by a representative from the state department of natural resources, who was unable to find evidence of gas in the slide area. The explosion hypothesis had some claim to plausibility, since the Associated Oil Company had encountered gas in exploratory wells drilled near by a few years ago.

Bedrock exposed by the slide consists of soft shale and siltstone, with minor amounts of sandstone and conglomerate. Much of the shale is exceedingly fine grained and except for surface drying remained wet and slippery two weeks after the slide occurred. No systematic search for fossils was made, but three species collected were identified by Dr. H. G. Schenck as the fresh-water forms *Amnicola longinqua* Gould, *Gonidea coalingensis* Arnold and *Fluminicola williamsi* (Hannibal), which identify the rocks as Upper Pliocene—correlatives of the Tulare formation in the San Joaquin Valley and of the Santa Clara formation farther north.

Further details about local geology were kindly supplied by Mr. F. A. Menken, chief geologist of the Associated Oil Company. Menken states that this part of the Lomerias Muertas is made up of interbedded marine and nonmarine strata on the southwest limb of an anticline. Although the San Andreas fault lies a scant 3 miles to the west, and a smaller fault parallels the slide ½ mile to the east, Menken believes that no significant faulting is associated with the slide area itself.

Although a predisposing cause for the slide is evident enough, just why it occurred at this particular time and place remains an unanswered question. The time seems especially odd: although the slide occurred during and after rains, the season as a whole had been exceptionally dry. No known seismic disturbance or artificial shock dislodged the slide as other masses in the vicinity were dislodged by the 1906 earthquake. Perhaps weathering and leaching by the rains of several years had prepared the material so that but one more wetting was needed; or perhaps, as suggested by Dr. Olaf P. Jenkins, dynamite blasts in the near-by oil wells had so deflected groundwater circulation that the bedrock was more than ordinarily moist

(in support of this is the presence of a small spring near the base of the wall exposed at the head of the slide).

GENERAL DESCRIPTION OF THE SLIDE

The disturbed area (Figs. 1 and 2) measures just under a mile in total length and almost 1,800 feet in greatest width. Movement was predominantly northward, from the main ridge of the Lomerias Muertas toward the Pajaro River. From the slide's foot to the top

FIG. 1.—General view of Gilroy slide, looking southeast. Main slide, with "cirque" and "moraines," in background. Toe of western part of slide in foreground. Older landslide scars on adjacent slopes. (All photos by A. C. Waters.)

of the cirque at its head is a difference in elevation of about 600 feet, so that the average gradient of the slope before the slide was about 1:8 (7°). The slope is considerably steeper in its upper part, the principal change in slope occurring just above (south of) square 4 in Figure 2.

The slide has two distinct parts: on the east, a jumbled mass of soil, vegetation, and rock fragments which has moved a considerable distance (hereafter called the "main slide"); on the west, a smaller area in which the movement has been relatively slight, so that the original sod is largely preserved except for numerous faults and gaping fissures. The predisposing cause of movement in both parts was saturation of mantle rock and the soft, shaly bedrock with water. Reasons for the difference in extent of movement are apparently the steeper slope in the upper part of the main slide and the deeper penetration of water into bedrock beneath it. Movements of the two

parts were independent, although one may have acted as a trigger to start the other. Debris from the main slide has overridden a portion of the western area.

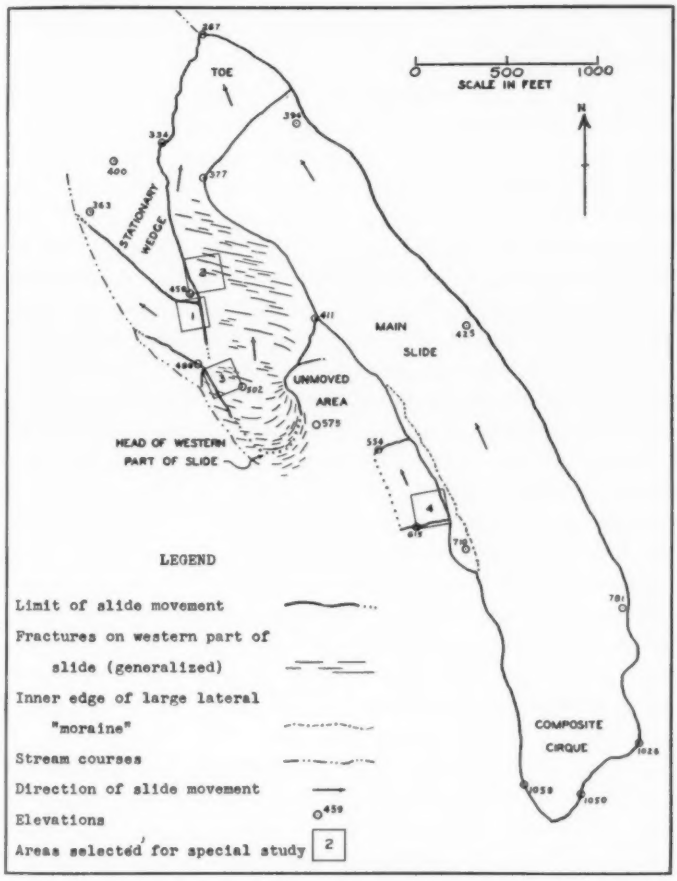

FIG. 2.—Outline map of the Gilroy slide

The main slide began movement as a huge slump, leaving at the top of the ridge a composite, steep-walled amphitheater (walls 200–300 feet high) reminiscent of a glacial cirque. Moving partly by slid-

ing, partly by flowage, its materials tumbled and crushed into incoherent debris, the mass invaded a small stream valley and followed it downward. Like a glacier, the slide piled up in its middle course two narrow ridges of debris resembling lateral moraines, with slickensided and striated inner surfaces (only the most conspicuous morainal ridge, on the west side near square 4, is shown in Fig. 2). The surface of the slide is rough and hummocky—in part from differential movement and in part from original topographic irregularities which were overridden. According to Sharpe's[1] classification of landslides and similar phenomena, the main slide has characteristics allying it with both landslides (*sensu strictu*) and debris avalanches.

The movement, distribution of material, and topographic forms produced in the main slide might well be a subject for detailed study in themselves, but the chief concern of this paper is with the western part of the slide. Here well-preserved structures give clear evidence of the forces brought into play in the early stages of landslide movement. Maximum movement in this part of the slide probably did not exceed 100 feet; at the only point where a quantitative estimate was found possible, a short distance north of square 2, the movement was about 80 feet. The surface of the slide is gashed and faulted, but only in a few places was the motion sufficient to fracture the sod and underlying material into a jumbled mass of debris. The present gently undulating topography of the slide is almost certainly a good replica of the original topography, albeit stretched and displaced somewhat downslope. Probably the movement was in part flowage taken up in mantle rock within 30–40 feet of the surface, in part a sliding on lubricated clay surfaces. In its main features this part of the slide resembles the forms which Sharpe calls "earthflows."

The western part of the slide (hereafter, for brevity, this part of the slide will be designated simply as "the slide") is bifurcated by a ridge called the "stationary wedge" on Figure 2. Evidently the wedge was too high for the slide to override it, and its materials insufficiently water-soaked to take part in the movement. The larger fork of the moving mass chose the eastern slope of the wedge,

[1] C. F. S. Sharpe, *Landslides and Related Phenomena* (New York: Columbia University Press, 1938).

sliding down into the valley of the main slide; the lesser fork moved toward the small valley west of the wedge.

The slide heads on a steep hillside gashed with concentric fissures and slumps. (See Fig. 2. Lines on the map indicate directions and general abundance of fractures but do not themselves represent individual fractures. Only on Figs. 5, 7, 8, and 10 are the details of individual fractures portrayed.) Downslope on either side of the slumped area the fissures become gash fractures against a strike-slip fault which marks the boundary of the movement. On the east the strike-slip fault is well defined even in its upper portion; it curves down a steep hillside, then follows a north-northeast course until it is buried by the "moraine" of the main slide. On the west the strike-slip movement is at first suggested only by gaping gash fractures, a definite surface of strike-slip faulting becoming apparent only near square 3. Beyond square 3 the boundary fault veers westward, rather oddly following the brink of the small stream valley rather than descending to the stream. Between square 3 and square 1 begins another strike-slip fault, trending a bit west of north, its beginnings lost in the maze of fissures just north of square 3; farther on, this becomes the boundary fault of the slide's main fork, east of the stationary wedge. From it in square 1 branches the small strike-slip fault which bounds the western fork.

Across the toe of the slide is a series of transverse, broken ridges, suggesting compression as friction brought the sliding mass to a stop. The margin of the toe is made up of a series of rounded bulges, its better-preserved portions showing recumbent folds piled one on the other, nappe fashion (Fig. 3). Unfortunately, differential movement within the slide has here been sufficient to fragment the sod and largely obliterate the structures.

Dominant on the surface of the slide are the effects of tension. Open fissures are abundant, and normal faulting is the rule. Especially conspicuous is the display of successive normal faults with their upthrow blocks to the north in the vicinity of square 2, ideal illustrations of the type of displacement called by Cloos[2] "antithetic faults" (faults whose direction of movement is opposed to the prin-

[2] H. Cloos, "Über antithetische Bewegungen," *Geol. Rundschau*, Vol. XIX (1928), p. 246.

cipal movement—here the downhill sliding) (Fig. 4). Graben structures are well exhibited at the tops of some of the small hills on the slide. The slumps at the head of the slide give good examples of "synthetic faults" (faults whose movement is in the same direction as principal displacement), although many of the slumped blocks have rotated forward sufficiently to show antithetic faulting as well. The prevalence of tension structures over the slide's surface suggests

FIG. 3.—Part of toe of slide, showing "nappe structure" made up of recumbent folds.

that the mass moved pretty much as a unit, stretching itself downslope as it moved.

The strike-slip faults at the edges, the slumps at the head and the overthrusts at the toe, and the numerous fractures on the surface of the slide display a wide variety of fault structures like those often associated with plastic flow. Here a thick surface layer of mantle rock and soil acted as the plastic medium, apparently a nearly homogeneous one, only the sod of the uppermost layer being somewhat stronger and more brittle than the rest. In the following section, structural types produced in this plastic medium under the influence of different forces and different original topographic forms will be analyzed by detailed examination of three selected areas (squares 1, 2, and 3 in Fig. 2). A fourth area is included (square 4) to show the effects of pressure in a similar medium.

DETAILED STUDY OF SELECTED AREAS

The outline map of Figure 2 was drawn from a rough survey with plane table and alidade. Distances should be accurate within 2 per cent and relative elevations within 4 per cent. Absolute elevations were obtained by comparison with the San Juan Bautista topographic map and may be as much as 50 feet in error. The four detail

FIG. 4.—General view of western part of slide, from just east of square 3. Car stands near tip of stationary wedge, just above large fault bounding eastern fork. Antithetic normal faults on surface of slide. Small hilltop grabens in foreground. Tip of main slide on extreme right. Pajaro Valley in background.

maps, Figures 5, 7, 8, and 10, were drawn from more careful surveys, using a small plane table, a Brunton compass with alidade attachment, and a steel tape. Some generalization was unavoidable, even with a scale of 20 feet to the inch, but the dimensions and locations of most of the structures shown are believed to be accurate.

Square 1, strike-slip structures (Fig. 5).—This area shows in detail the splitting of the slide at the tip of the "stationary wedge." The major line of strike-slip movement bounding the east fork of the slide crosses the area in an approximately north-south direction, and from it branches the minor strike-slip fault along which the small western

fork moved. These intersecting faults divide the area into three major parts: the stationary wedge on the north, the badly fractured "eastern block," and the nearly unfractured "western block." Movement of the eastern block to the north in this area probably averaged between 20 and 30 feet, and movement of the western block to the

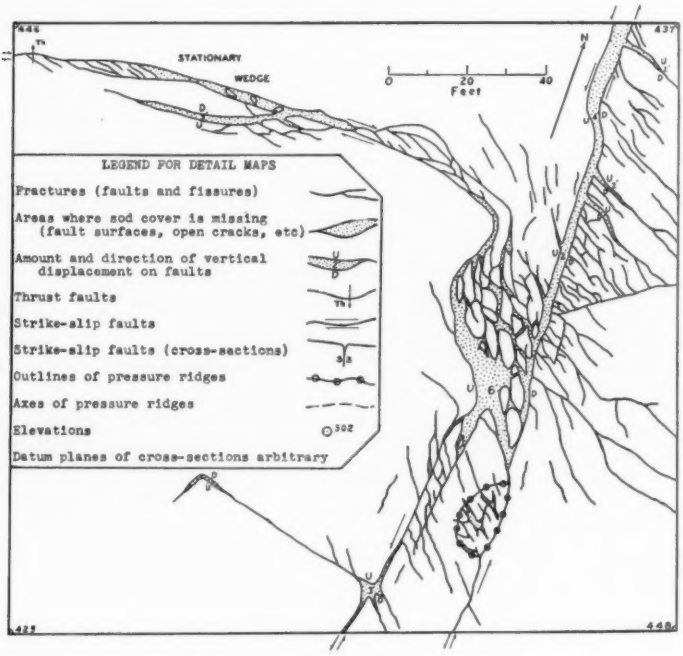

FIG. 5.—Detail map of square 1

west was probably between 5 and 10 feet. Topographically the stationary wedge is higher than adjacent parts of the other blocks. The eastern block is downdropped 2–6 feet along the fault line and slopes gradually eastward; the western block has a saddle just south of the stationary wedge, then rises toward the south boundary of the area.

Features commonly associated with strike-slip movement are evident from the map (see also Fig. 6). The faults for the most part are zones of faulting rather than single surfaces. Secondary faults deploy

from the main fractures, making acute angles with the direction of movement. Along the smaller strike-slip fault these secondary breaks frequently show arcuate patterns, some of them outlining sliver blocks. Small blocks in the zone of movement are rotated. The small overthrust in the northwest corner is associated with a sharp turn in the strike-slip movement, the fault curving around the stationary wedge to the northwest just beyond this area (see Fig. 2).

FIG. 6.—Strike-slip fault bounding western block, just west of square 1. Gash fractures and sliver blocks along fault.

Movement of the western block requires its pulling away from the eastern block along the line of major strike-slip movement. The necessary stretching is taken up in part by the system of fractures in the southern part of the area. Effects of stretching are particularly noticeable in the much-disturbed zone at the very tip of the stationary wedge.

The eastern block has moved downward to the east as well as northward with relation to the other blocks, so that the major strike-slip fault is a line of normal faulting also; toward the south edge of

this area the normal faulting dies out in a fractured monocline. Probably the numerous small cracks and faults on the eastern block trending roughly N. 60° W. are tension fractures produced by the northward and eastward movement and have no direct connection with the strike-slip fault.

In contrast to the obvious tension along the main fault, evidence points to compression between the western block and the stationary wedge. Here the sliding material was moving down against the head of the wedge, attempting to carry the latter with it. Perhaps the greater friction along this fault accounts for its conspicuously different pattern of secondary fractures. Whereas minor fractures along the main north-south fault branch from it at large angles and are fairly straight, secondary fractures here branch at low angles and are often curved. Elongated sliver blocks between curved fractures are striking features along this fault but are absent along the main fault. That patterns of arcuate, low-angle secondary faults and elongated sliver blocks are characteristic of strike-slip movements involving compression is also suggested by the strike-slip fault of Figure 8, where compression is shown by the overthrust to the northwest (see p. 645). By contrast these features are practically absent along the large strike-slip fault mapped in Figure 7 (another section of the north-south fault of this area), where tension was dominant.

Square 2, normal (antithetic) faults (Fig. 7).—This area, a short distance north of square 1, was chosen to show the relation to the western fault of the normal faults which are so prominent a feature of the slide's surface. The area includes a narrow strip of the stationary wedge along its west margin and a larger segment of the much-fractured eastern block. The surface of the eastern block slopes gradually to the northeast, as the cross section shows (see also Fig. 4).

The strike-slip fault separating the two blocks is here a wide zone of disturbance in which few structures are visible. Along this fault the eastern block has dropped about 8 feet and moved northward probably 40–50 feet; horizontal striations on the exposed fault surface furnish a clear record of the northward movement. A few minor strike-slip faults parallel the major one, and some of the small branching cracks are probably gash fractures produced by the strike-

slip movement. Reasons for the angles made by some of these secondary fractures with the main fault are not clear. The single small pressure ridge is probably explained by the adjacent bulge in the strike-slip fault.

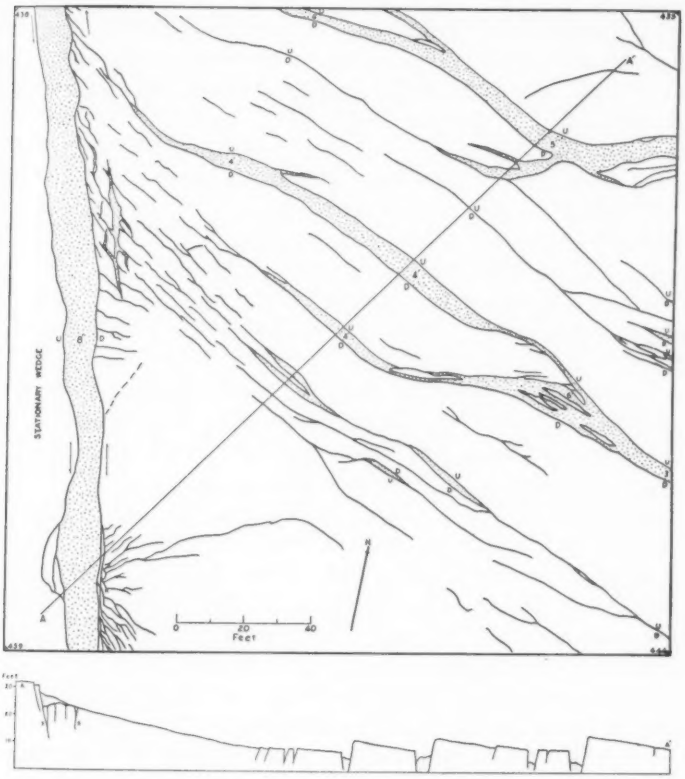

FIG. 7.—Detail map of square 2 and cross section

The fractures on the eastern block are nearly all open gashes, the larger ones showing vertical displacements of small magnitude (up to 5 feet). With a few minor exceptions the upthrow sides of these faults are on the north. Tension is obviously responsible for this system of open fissures and the normal faulting may be explained by

the attempted forward rotation of parts of the moving mass owing to friction at its base.

The orientation of these tensional fractures seems at first sight a bit peculiar. They are conspicuously parallel, their average strike remaining about N. 60° W. over a considerable area outside the mapped square. Now the slide in this part of its course was evidently moving nearly due north; seemingly the maximum tension should be along the line of movement and tensional fractures should accordingly strike east-west. Probably topography may be held accountable for the oblique angle: the eastern block slopes eastward, toward the valley of the main slide, the slope increasing beyond the limits of this area. Hence the slide was here moving across an incline and was stretched not only in the direction of its length but also by the tendency of its material to slide downslope eastward. A combination of tension in the direction of principal motion and tension toward the east gives the resultant tension in the direction N. 30° E. inferred from the fractures.

Square 3, graben structures (Fig. 8).—The original topography of the slide's upper portion was characterized by small rounded hills and valleys. Movement of the slide has produced gaping tensional fractures on the hills and small pressure ridges in the valleys. On the larger hills slight vertical displacements along the tension cracks produce rude graben-like structures, of which this area shows the best example. Typically the center of the graben is slightly north of the ridge of the hill and the antithetic faults on the north side of the graben are more numerous and better developed than the synthetic faults on the south side.

Several explanations for the graben-like structures suggest themselves but perhaps the simplest is the following: Tension cracks develop on hilltops because the uppermost layer of the slide is stretched abnormally in accommodating itself to motion over a curve which is convex upward (and compression ridges develop in valleys because here the surface layer must accommodate itself to a curve which is concave upward). The largest of the tension cracks will be near the crest of the hill, where the curvature is greatest. These tension fractures divide the hill into blocks which thereafter move with some measure of independence. As the general sliding movement con-

tinues, dragging the blocks with it, blocks on the downslope (north) side of the hill tend to fall forward, producing antithetic faults, while those on the upslope side of the hill fall backward, producing synthetic faults. Since the downslope side is commonly the steeper, the

FIG. 8.—Detail map of square 3 and cross section

antithetic faults are better developed and have lower angles than the synthetic faults.

In this area the hill is roughly T-shaped, the crossbar of the T lying along the region of greatest disturbance (about N. 70° W.), the vertical line of the T extending approximately along the line of

cross section BB' to the southwest corner of the map. The steepest slope is toward the north. A small valley lies along the south border of the map, turning to the north at the southeast corner. These topographic relations are well shown in Figure 9.

The graben structure is here complicated by a strike-slip fault cutting across its western end. This is the western boundary fault, just south of the point where it turns abruptly westward (see Fig. 9). Along the line of section BB' it happens that the western side of the

FIG. 9.—Large graben, square 3. Looking west along center of graben. Cliff on left in part represents line of strike-slip movement.

strike-slip fault is upthrown relative to the eastern side. This adds an apparent synthetic fault to the graben structure and makes the graben look more perfect than it is. The other synthetic faults shown on the cross section are very small, as might be expected, since the hill here has little slope to the south.

The center of the graben is a broad chasm 8–12 feet deep, with a jumbled mass of soil and bits of sod in its bottom and a few isolated sod-covered blocks still standing. The very numerous antithetic faults to the north divide the sod into narrow slivers which lie one against the other. To the east the graben disappears in the small valley mentioned previously; to the west it ends against the strike-slip fault in a region of great disturbance.

The average strike of the graben faults is approximately the same as that of the normal faults farther down on the slide. Here the direction is probably governed chiefly by topography. The major movement of the slide in this area, as suggested by the strike-slip fault, is approximately N. 30° W., but the maximum tension over the hill is determined by the steepest slope, which lies approximately along the line of cross section (about N. 15° E.). The tension cracks are developed transverse to this direction. Topographic influence is also suggested in the orientation of the pressure ridges in the small valley along the south border, since axes of these ridges lie approximately parallel to the axis of the valley.

The strike-slip fault which crosses the western part of this area begins a short distance to the south in a maze of open gashes like those shown in the southwest corner of the map. All across this area the strike-slip movement is well defined, with the characteristic sliver-block and gash-fracture features noted in square 1. That there has been considerable pressure across the fault is shown by a slight overthrusting of the surface material west of the fault. This overthrusting appears on the map (Fig. 8) in the northwest corner.

The zone of disturbance at the north border of the map, where the graben meets the strike-slip fault, continues for some distance to the north. From this zone of disturbance springs the major strike-slip movement shown in squares 1 and 2.

Square 4, compression structures (Fig. 10).—The best locality for studying structures due to compression is this area at the western border of the main slide, where its "moraine" is largest. Here, just below the main slide's steepest portion, debris from the cirque above pushed with considerable force against the west bank and in part overrode it. One result of the compression was the loosening of a rectangular block about 400 feet wide and 600 feet long (Fig. 2). The block moved a few feet downslope, producing a large normal fault at its southern end and several small overthrusts at its northern end. Square 4 includes the southeast corner of this block.

The principal effect of pressure on the surface of the moved block is the development of anticlinal ridges, most of them with their axial planes tilted westward. These ridges are well shown in Figure 11, which includes approximately the same area as the map. On the

map, outlines of only the larger ridges are shown; smaller ones are indicated simply by dashed lines along their axes. The original topography here was evidently a gentle and approximately uniform slope from southeast to northwest.

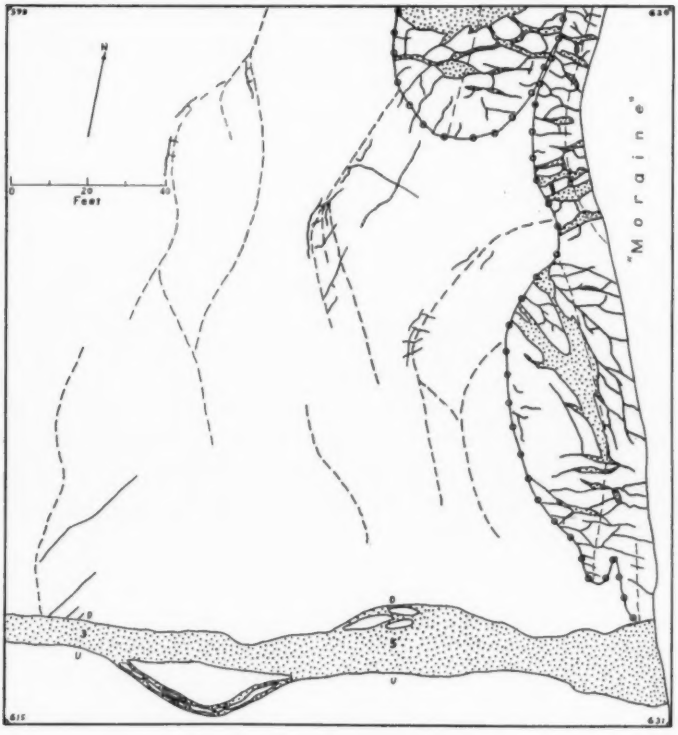

FIG. 10.—Detail map of square 4

The general pattern of anticlinal ridges shows the effect of pressure in an approximately east-west direction. Although the larger folds are badly broken and are cut by faults in apparently random directions, two general types of fracture are discernible: longitudinal gashes where the anticlines have broken along their axes and cross fractures approximately at right angles to the axes. The latter are

in part shear faults produced by differential movement in the folds, corresponding to the so-called "flaws" that develop in the upper plates of thrusts, but the majority are apparently simply tensional gashes produced by the bowing-out of the folds to the west.

COMPARISONS WITH SIMILAR STRUCTURES

Structure patterns of this sort invite comparison with similar structures produced on a smaller scale in laboratory clay boxes and

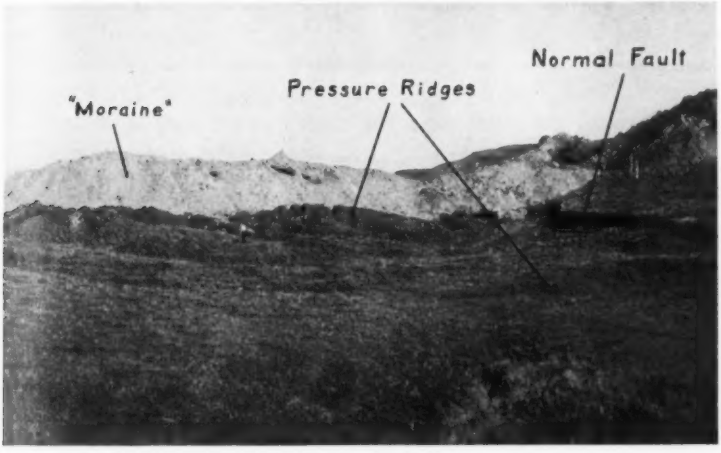

FIG. 11.—"Moraine," pressure ridges, and fault, square 4. Looking east at southeast corner of moved block.

on a larger scale in rocks. Features associated with the strike-slip faults of square 1, for instance, are similar even in small details to structures mapped along large strike-slip faults, and also to patterns produced by strike-slip movements in laboratory experiments. The antithetic normal faults of square 2 closely resemble faults produced by tension in the clay-box experiments of Hans Cloos[3] and suggest a type of structure frequently proposed for such areas as the Great Basin. The anticlinal ridges of square 4 are similar in origin and show similar fracture patterns to some of their larger counterparts in

[3] "Künstliche Gebirge," *Nat. u. Mus. Senckenbergische Naturforschende Gesellschaft* (Frankfort am Main, 1929–30), Part I, pp. 225–43; Part II, pp. 258–69.

solid rocks. Graben structures like those of square 3, if the proposed mechanism for their formation is correct, are probably not duplicated in rock deformation, nor so far as we know have grabens been produced by a similar mechanism in experiments.

To estimate quantitatively the validity of such comparisons would require a knowledge of three things: the forces involved in producing the structures, the time involved, and the strength of the material.[4] Reasonable guesses for the first two of these quantities could easily be made, but the third remains elusive because no record is available as to just how wet the slide material was when it moved. In the absence of this datum the extrapolation from these structures to other structures on a larger or smaller scale becomes dangerous. Comparisons are suggestive, but their validity cannot be established.

ACKNOWLEDGMENTS.—We are indebted to Dr. A. C. Waters for suggesting the problem, for permitting us to use several photographs, and for many helpful comments in the course of the work. Dr. Olaf P. Jenkins discussed the slide area with us and offered some valuable suggestions. Mr. F. A. Menken and Dr. H. G. Schenck kindly gave us information about the geology of the region. To Mr. J. W. Thompson, manager of the San Benito Ranch, we are indebted for help in the field work.

[4] M. K. Hubbert, "Theory of Scale Models as Applied to the Study of Geologic Structures," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 1459–1520.

THE STATUS OF *CEPHALASPIS SCHRENCKII* PANDER FROM THE UPPER SILURIAN OF OESEL

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ABSTRACT

Cephalaspis schrenckii Pander is a Silurian ostracoderm generically distinct from both *Cephalaspis* and the forms to which the term *Trematapis* is usually applied. Under strict application of taxonomic law this species is the genotype of *Trematapis*. This would result in confusion; it is advocated that the rules be suspended and *C. schrenckii* be made the type of a new genus, *Witaaspis*.

In 1856 Pander[1] published a remarkable paper on the fossil fishes of the Silurian system of the Russian-Baltic provinces, in which he described twenty-eight genera and forty-two species. Many of his species were founded upon isolated coelolepid scales or upon small fragments of the exoskeleton of other ostracoderms. How many of his genera and species would remain valid today is as yet an unanswered question.

One species founded on incomplete specimens was named *Cephalaspis schrenckii*. Ten years later Schmidt[2] established the genus *Trematapis* with one species. He united Pander's *Cephalaspis schrenckii* with this new material and gave the species the name *Trematapis schrenckii*.

In 1892 Rohon[3] gave a comprehensive description of *Trematapis*, separating Schmidt's material from *Trematapis schrenckii* under the name *Trematapis schmidti*, and founding two additional species: *Trematapis mickwitzi* and *Trematapis simonsoni*.

[1] C. H. Pander, *Monographie der fossilen Fische des silurischen Systems des russisch-baltischen Gouvernments* (St. Petersburg: Buchdr. der K. Akademie der Wissenschaften, 1856).

[2] F. Schmidt, "Über *Thyestes verrucosus* Eichwald und *Cephalaspis schrenkii* Pander, nebst einer Einleitung über das Vorkommen silurischer Fischreste auf der Insel Oesel," *Verh. Russ. Min. Ges.* Vol. I (2d ser., 1866).

[3] J. V. Rohon, "Die obersilurischen Fische von Oesel, Part I: *Thyestidae* und *Trematapidae*," *Mem Acad. Imp. Sci. St. Pétersbourg*, Vol. XXXVIII, No. 13 (7th ser., 1892).

In 1893 Schmidt[4] published a brief note recognizing Rohon's separation of *Trematas pis schmidti* from *Trematas pis schrenckii* and restoring the latter to its original designation of *Cephalaspis*. He united Rohon's *Trematas pis simonsoni* with it.

By 1894 Schmidt had secured further material, including specimens showing an occipital region with traces of segmentation somewhat similar to those found in *Thyestes*, and he transferred it once more—this time to *Thyestes schrenckii*.[5]

In 1894 Rohon[6] published an appendix to his 1892 paper, giving a further description of *Trematas pis schmidti* and removing *Trematas pis schrenckii* from the genus, though not accepting Schmidt's inclusion of *Trematas pis simonsoni*.

In 1895 Rohon[7] discussed the segmentation of the head in the Thyestidae, and much of his information was obtained from a specimen listed as *Thyestes schrenckii*, apparently accepting Schmidt's placement of the species.

The question seems to have rested there until Stensiö's monographs on the cephalaspids of Spitzbergen and of Great Britain. In the former[8] he stated that it was impossible to decide with certainty whether the species belonged to *Thyestes* or represented a genus of its own, since it was as yet too imperfectly known. In the latter[9] he states that this species does not belong to *Thyestes* but probably represents a genus of its own.

Analysis of the nomenclature of this species reveals an error of long standing which was called to my attention by Dr. A. S. Romer after he had read a first draft of this paper.

[4] "Über neue silurische Fischfunde auf Oesel," *Neues Jahrb. für Mineralogie*, Vol. I (1893).

[5] "Über *Cephalaspis* (*Thyestes*) *schrencki* Pander aus dem Obersilur von Rotziküll auf Oesel," *Geol. et Pal. Bull. Acad. Imp. Sci. St. Pétersbourg*, Vol. I (1894).

[6] "Zur Kenntniss der Trematas piden," *Geol. et Pal. Bull. Acad. Imp. Sci. St. Pétersbourg*, Vol. I (1894).

[7] "Die Segmentirung am Primordial-cranium der obersilurischen Thyestiden," *Verh. Russ. Kais. Min. Ges.*, Vol. XXXIII (1895).

[8] E. A. Stensiö, "The Downtownian and Devonian Vertebrates of Spitzbergen, Part I: Family *Cephalaspidae*," *Skrifter om Svalbard og Nordishavet*, No. 12 (1927).

[9] *Cephalaspids of Great Britain* (London: British Museum Natural History, 1932).

A genus is not founded on a specimen or on specimens, but on a species, and the genus thus founded follows that species in whatever its fate may be. Thus when Schmidt[10] established the genus *Trematalaspis* he took *Cephalaspis schrenckii* Pander as the genotype. As a matter of fact, Schmidt at that time was unable to find the specimens on which Pander had based his description, and among the other material collected from Oesel he found only one specimen showing the six-sided fields which Pander had given as a characteristic feature. He regarded these fields as similar to the polygonal areas seen through the polished superficial layer of *Trematalaspis* and apparently believed them to be specimens from which this layer was missing. The fact that his description was based on specimens which he mistakenly regarded as conspecific with Pander's material does not alter the taxonomic situation. Pander's species is the genotype of *Trematalaspis;* the material which Schmidt described and which was later split off as *Trematalaspis schmidti* is not.

By 1892 Rohon had discovered three specimens of the Pander species. When Schmidt[11] restored Pander's species to *Cephalaspis schrenckii* he ought to have given a new generic name to "*Trematalaspis*" *schmidti*—the name *Trematalaspis* becoming a synonym for *Cephalaspis*—as long as Pander's species remained in that genus, and later for *Thyestes* when the species was again transferred.

During my examination of the material collected by the late Dr. William Patten of Dartmouth College I have found a suite of specimens from the Wita quarry, collected in 1930, among which are fifty-seven specimens of the Pander species. My study of this material indicates that it does not fit into any of the present genera.

If taxonomic procedure were to be followed I ought, then, to restore the generic name *Trematalaspis* to this, the original genotype. To redescribe Pander's *Cephalaspis schrenckii* as *Trematalaspis schrenckii*, and to change the family and genus of what we have known for so many years as *Trematalaspis* would, however, seem to produce greater confusion. Thus I am asking for suspension of rules in this case and am giving the Pander species a new generic name, *Witaaspis*, from the quarry in which the specimens have been found.

[10] "Über *Thyestes verrucosus* Eichwald . . . ."

[11] "Über neue silurische Fischfunde auf Oesel."

Cephalaspidae

Witaaspis N. GEN.

This genus includes cephalaspids of small size with rudimentary cornua somewhat like those of *Hemicyclaspis* (Fig. 1). Pectoral sinuses are rudimentary or absent. The trunk division of the shield is of moderate length, narrowing at the level of the cornua to about three-fourths the width of the head shield at that level. No certain traces of segmentation occur on the posterior portion of the dorsal shield, although in one specimen the transverse boundaries of a number of adjacent tesserae are aligned, giving a semblance of segmentation.

The anterior part of the shield slopes sharply from the nasal eminence. The orbits are about one-third of the shield length from the anterior margin. The dorsal field is large, oval, and bordered laterally by ridges which unite behind it to form a median crista. The lateral fields are single. Their posterior extent is uncertain, but seems to have been nearly to the cornua. The shield is thin.

Witaaspis schrenckii (PANDER)

1856 *Cephalaspis schrenckii* Pander (*op. cit.*, Pl. IV, Fig. 2, p. 47).
1866 *Trematalaspis schrencki* Schmidt (in part) ("Über *Thyestes verrucosus* . . . .").
1892 *Trematalaspis schrencki* Rohon ("Die obersilurischen Fische von Oesel," Pl. II, Figs. 14 and 15, p. 61).
1893 *Cephalaspis schrencki* Schmidt ("Über neue silurische Fischfunde auf Oesel," p. 99).
1894 *Thyestes schrencki* Schmidt (in part) ("Über *Cephalaspis* [*Thyestes*] *schrencki* Pander . . . . ," Pl. I, Figs. 3, 4, 5, 6, 9, ?7, ?8 [not Figs. 1 and 2], p. 206).
1895 *Thyestes schrencki* Rohon ("Die Segmentirung . . . . ," Figs. 1–5).
1927 "*Cephalaspis*" *schrencki* Stensiö ("The Downtownian and Devonian Vertebrates" [not Pl. XLVIII], Fig. 2, p. 295).

Witaaspis schrenckii is a small cephalaspid. The length of the shield from the rostral end to the posterior end of the crista is about 15–17 mm.; the width at the level of the pectoral sinuses is about 20 mm. The dorsal shield is more abruptly arched anteriorly than in most *Osteostraci* (Fig. 2). Anteriorly and laterally it has a raised, marginal rim, about 1.25 mm. wide, going smoothly over the edge. The part of the rim which goes over onto the venter is ornamented

with tubercles, slightly elongated, parallel to the margin. The granular ornamentation of the more dorsal parts of the rim is not so definitely arranged in fields as it is over the general dorsal surface, but more nearly resembles that found on *Rotsikŭllaspis obrutchevi* Robertson.[12] Over the dorsal surface the ornamentation consists of fine granular tubercles which appear under the microscope like fine sand grains ornamenting slightly arched polygonal areas, frequently six-sided, separated from each other by smooth, shallow furrows

FIG. 1

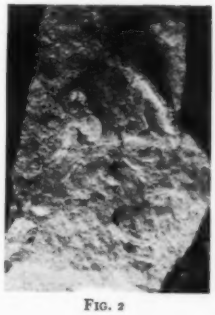

FIG. 2

FIG. 1.—Reconstruction of *Witaaspis schrenckii*. ×*ca.* 3

FIG. 2.—*Witaaspis schrenckii*. Specimen c19, Patten Collection. × *ca.* 3

(Fig. 3). The cornua and pectoral sinuses are rudimentary (Figs. 1 and 4). The trunk division of the shield is moderately long. The orbits are nearly circular, their longitudinal axis slightly longer than their transverse axis, and large in size. They are slightly tipped outward anteriorly. The large nasal fossa is nearly circular in outline—the outline being slightly altered both anteriorly and posteriorly by a prominent median ridge which carries the large, keyhole-shaped aperture (Fig. 2). One specimen shows granular ornamentation in the floor of the fossa. The pineal foramen is incompletely preserved but appears to have been large and nearly circular. No trace of the pineal plate or of ossified sclerotics has been found. The large oval dorsal field is bounded laterally by ridges which converge to join the crista at the posterior end of the field. This crista is relatively low.

[12] G. M. Robertson, "New Genera of Ostracoderms from the Upper Silurian of Oesel," *Jour. Paleon.*, Vol. XII (1938), p. 490.

The lateral fields have been described above. Traces of some portions of the sensory canal system are preserved. The infraorbital runs a course somewhat like that of *Dartmuthia* and *Trematas pis*.[13] It starts lateral to the posterior end of the orbit (Fig. 1), runs anteriorly and slightly laterally along the border of the orbit, then turns more laterally and runs anterolaterally to near the level of the anterior end of the lateral field. Here it turns medially and slightly anteriorly to end near its fellow in the mid-line. Traces of what I have termed the postorbital[14] occur on a few specimens, directed

FIG. 3

FIG. 4

FIG. 3.—*Witaaspis schrenckii*. Specimen *c*16, Patten Collection. × *ca*. 3

FIG. 4.—*Witaaspis schrenckii*. Specimen *c*28, Patten Collection. × *ca*. 3

posteriorly and slightly laterally from a point posterolateral to the end of the infraorbital canal. In one specimen there appears to be a faint line posterior to this, starting lateral to the posterior end of the dorsal field. This may be the anterior end of the main lateral line. In one specimen two faint lines are found running transversely from near the posterior end of the dorsal field in about the position of the supratemporal line.[15] The only ventral features known are a few traces of the anterior part of the rim. There was in the median part of the anterior margin a slightly concave area, much less marked than that of *Trematas pis*—rather broad and shallow. This probably

[13] Robertson, "The *Trematas pidae*," *Amer. Jour. Sci.*, Vol. XXXV (1938).

[14] *Ibid.* [15] *Ibid.*

indicates the position of the mouth, which I regard as having been suctorial.

Some portions of the endoskeleton are preserved both in deeply eroded dorsal aspects and in ventral aspects of the dorsal shield. The

FIG. 5.—*Witaaspis schrenckii*. Specimen *c*42, part and counterpart, Patten Collection. × *ca*. 4.

branchial tubes and interbranchial ridges are well shown in figures published by Rohon.[16] Figure 5 of the present paper shows these, as well as the postbranchial wall and traces of otic vestibules. In these features *Witaaspis* is very cephalaspid-like. The ventral aspect of

[16] "Die Segmentirung. . . . ."

nasohypophyseal tube and the hypophyseal fossa (Fig. 6) are similar to Stensiö's Plate 48, Figure 2 ("The Downtownian and Devonian Vertebrates . . . ."). The hypophyseal fossa is broad and shallow, the tube being slightly deeper at its deepest part than is the fossa.

Certain of the nerve channels to the lateral fields are partly preserved in some specimens (Figs. 5 and 6). I am not positive of the most anterior, but there seem to have been at least five, and probably six, nerves. Unfortunately they are not preserved centrally.

FIG. 6.—*Witaaspis schrenckii*. Specimen *c*49, Patten Collection. ×2

The form of the brain case is very incompletely shown. Specimen *c*28 indicates a prominent metencephalon and a mesencephalon probably telescoped with the diencephalon as in *Trematas pis*.[17]

This description does not fit all the specimens described and figured as pertaining to the species. Schmidt[18] described traces of two or three segments fused in the posterior part of the shield. As indicated above, I do not find any definite traces of such segmentation, except that the transverse furrows separating adjacent polygonal areas are sometimes aligned. Schmidt's Figure 1, Plate I, shows a specimen with definite segmentation—as definite as in *Thyestes verrucosus*. In the figure no trace is shown of the ornamentation characteristic of *Witaaspis*, but instead there are apparently scattered tubercles of larger size. I regard this specimen as pertain-

[17] Robertson, "The *Trematas pidae*."

[18] "Über *Cephalas pis* (*Thyestes*) *schrencki* Pander. . . . ."

ing to another species, though I am not prepared to identify it with anything as yet described.

Schmidt also described short but well-defined cornua, and his Figure 2 shows one cornu which resembles those figured for *Thyestes salteri*. This specimen also I regard as not pertaining to *Witaaspis*. Comparison of that figure with the very typical specimen in his Figure 9 seems to eliminate it. Schmidt himself questioned whether the specimens of his Figures 7 and 8 pertained to this species, and of these I am uncertain. Thus I would accept his Figures 3, 4, 5, 6, and 9, rejecting Figures 1 and 2 and questioning Figures 7 and 8.

Stensiö[19] figured a specimen, Plate 48, Figure 2, as the "so-called *Cephalaspis schrencki*"—a specimen showing one cornu similar to that shown in Schmidt's Figure 2. This specimen also shows the ventral aspect of one orbit, the hypophyseal sac and fossa, and some vessels. This specimen I also reject, regarding it as probably the same species as that represented by Schmidt's Figures 1 and 2.

[19] "The Downtownian and Devonian Vertebrates. . . . ."

LAYERING IN YELLOWSTONE RHYOLITE

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ABSTRACT

The layered structure of much of the rhyolite of Yellowstone National Park consists of either a wavy lamination or a coarse stratification. Iddings attributed the layering to spreading of heterogeneous lava. Evidence is herein presented in favor of that hypothesis. Brouwer has suggested that the development of some of the coarser stratification was influenced by crystallization along planes developed by shear when the lava was in a highly viscous state and no longer able to spread. The writer suggests an additional possibility—that the crystallization of some of the layers may have been induced by vapors not originally included within the lava, but which invaded pervious bands originally determined by flowage.

THE PROBLEM

The rhyolites of Yellowstone National Park commonly display a layered structure. J. P. Iddings,[1] after exhaustive megascopic and microscopic study, decided that the layering was due to the "spreading out through flowage of a more or less heterogeneous viscous fluid." H. A. Brouwer[2] has recently suggested that much of the layering in the rhyolite resulted from the development of planes of crystallization by shear, with the crystallization probably favored by circulation of gases. The shear planes are believed to have been developed by movements after the flow structure was completed, while the lava was in a highly viscous state. The present paper presents additional information on the layering of the rhyolite and proposes a variation of Iddings' hypothesis as a possible alternative explanation of some of the coarse layering.

LAYERED STRUCTURE

The layering consists of either a streaky, discontinuous, wavy lamination or a coarse stratification with layers from less than an inch to several feet thick. There are many variations of these two principal types.

[1] "Geology of the Yellowstone National Park," *U.S. Geol. Surv. Mono. 32*, Part II (1899), chap. x, pp. 356–432.

[2] "On the Structure of the Rhyolites in Yellowstone Park," *Jour. Geol.*, Vol. XLIV (1936), pp. 940–49.

The wavy lamination is found in much of the vitreous rhyolite as well as in the more crystalline facies. In thin section the lamination manifests itself in well-developed fluxion structure (Fig. 1). The varicolored streaks swirl around phenocrysts and other obstacles; and enterolithic fragments of earlier-formed glass are strung out in lines. The streaky lamination can be interpreted only as a direct result of flowage.

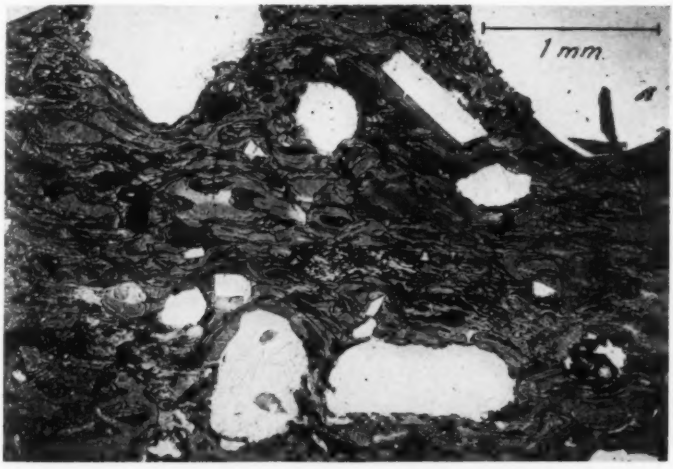

FIG. 1.—Fluxion structure in laminated rhyolite

The origin of the coarse layering, of which there are several types, is less obvious. The coarse layering referred to by Brouwer, and which is the subject of the present paper, consists of alternating beds of black vitrophyre (Fig. 2) and dense, reddish, tridymite-saturated rhyolite (Fig. 3). Because of their greater resistance, the red layers project as sharp ledges. In the vicinity of Lower Falls they maintain a uniform thickness for more than a score of feet.

THEORIES OF LAYERING

According to Iddings, all the layering of the Yellowstone rhyolite resulted from the flowage of a heterogeneous viscous fluid. He attributed the heterogeneity of the lava to the unequal distribution of

water vapor absorbed near the earth's surface by the rising magma. He believed that layering and localized crystallization were more common in acid lavas because of their greater viscosity. The viscosity is thought to have retarded diffusion and to have hindered an even distribution of the constituents of the lava.

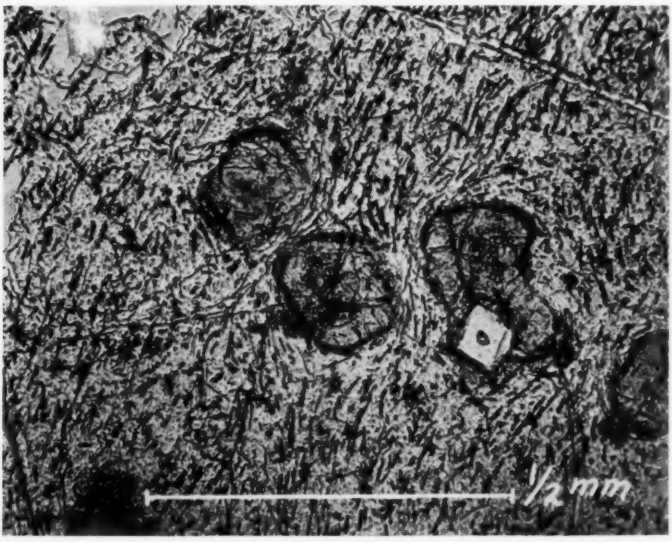

FIG. 2.—Microlites swirling around spherulites in vitrophyre a few feet upstream from brink of Lower Falls. The spherulites formed while the lava was still liquid. Uncrossed nicols.

The greater density of the reddish layers is due to a greater degree of crystallinity (Fig. 3). It was presumably Iddings' idea that portions of the magma rich in volatile substances were spread out in layers by flowage and that the volatiles stimulated the crystallization of these particular layers. Brouwer, however, believes that "much of the banding or lamination in Yellowstone Park is the result of crystallization in a late stage of solidification, when the development of the flow structure was completed and the lava was no longer able to spread."[3] He suggests that the crystallization "may have

[3] *Ibid.*, p. 943.

been greatly favored by circulation of gases along planes developed by movements when the lava had nearly come to rest."[4] Brouwer clearly recognized, however, that the lava did not everywhere yield to continued pressure by shearing, for, in discussing the larger folds in the rhyolite, he says: "Resistance in front and below has caused

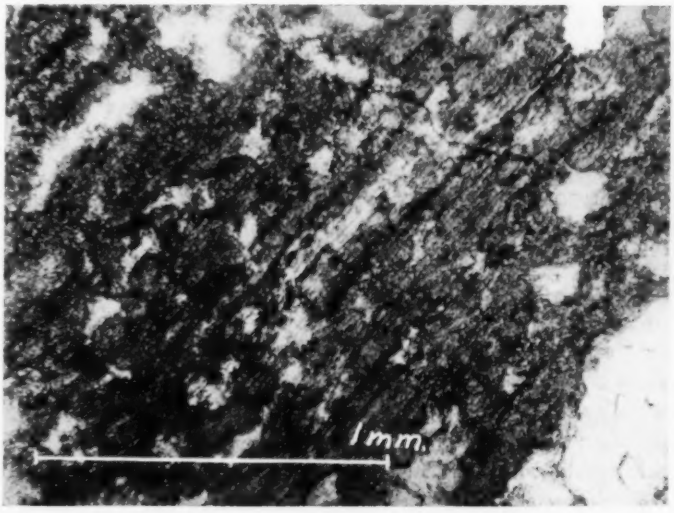

FIG. 3.—Flow-banded rhyolite showing selective penetration by tridymite (practically all the colorless patches are tridymite). Note also the crystallization of tridymite along the cross-cutting fracture. The groundmass is devitrified glass. Dense, reddish rhyolite at brink of Lower Falls. Uncrossed nicols.

bending in the still-moving mass, where conditions have favored this kind of yielding."[5] Whether he means to imply that shearing took place after movement ceased is not clear. C. N. Fenner[6] came to the same conclusion regarding folding of some of the rhyolites. He writes:

Their structure, and especially the banding, seemed to me to show that many of the flows moved as viscous fluids, that gradually became more and more stiff, until movement finally ceased. As a consequence of this mode of pro-

[4] *Ibid.*, p. 943.

[5] *Ibid.*, p. 945.

[6] Personal communication.

gression, steep faces of rigid material were sometimes formed, that were overflowed by more mobile liquid from behind, and this, in turn, congealed as it flowed over the dam.

Inasmuch as Brouwer recognizes that folding rather than shearing may take place under certain conditions, and inasmuch as precisely the same type of layering is found in the folded as in the unfolded rhyolite, it seems possible that one explanation may account for the layering in both instances. Consolidation of a heterogeneous lava seems to explain reasonably the layering both where the lava is undisturbed and where it is folded, whereas the shearing hypothesis seems to require a separate explanation for the layering in those folded areas where no shearing is believed to have occurred.

With regard to the spherulites found in much of the layered lava, Brouwer states that the "late stage in which the spherulites were formed is illustrated in many of the rocks when studied under the microscope."[7] He describes a layer of obsidian between Upper and Lower Falls in which the spherulites "are of later origin" than the flow structure. The occurrence of spherulites in layers, and their frequent elongation, is attributed to development along planes of crystallization formed by shear. In support of such late crystallization of the spherulites he cites Iddings' observation of spherulitic crystallization subsequent to cracking of the rhyolite, and Iddings' conclusion that the "magma was so viscous previous to its last movement that a small gap in it was not closed." Iddings' observation was quite correct: the development of spherulites along cracks is not uncommon in the rhyolites of Yellowstone. It is questionable whether the fact leads inevitably to the conclusion that movements continued after the development of the spherulites, but this matter is not of prime importance in the present discussion. It is important, however, that many spherulites do not have such a late date of origin. Figure 2, reproduced from an earlier publication,[8] shows a vitrophyre collected from one of the areas described by Brouwer. The swirling of the microlites around the spherulites indicates clearly that these spherulites formed while the lava was still

[7] *Op. cit.*, p. 944.

[8] R. J. Colony and A. D. Howard, "Observations on Spherulites," *Amer. Min.*, Vol. XIX (1934), pp. 515–24.

liquid, before conditions conducive to shearing had been attained. On the other hand, the writer has examined rocks in other localities where the flow lines pass undisturbed through the spherulites and has published reasons[9] for believing that this situation indicates a later age for the spherulites. The growth of spherulites around perlitic or other cracks also indicates that the spherulites grew after flow was completed, during contraction of the solid lava mass. Hence, there is evidence that spherulites and similar bodies may be either older or younger than flow structure. Iddings likewise found this to be true, for in discussing the vertically layered rhyolite at Natural Bridge on the west side of Yellowstone Lake he says:

> The lithoidal rock alternates with glassy layers of black perlite having dense spherulitic bands and some larger dense spherulites. . . . . The large hollow spherulites have been crushed while the matrix was plastic though not liquid, for the broken shells have been dislocated and the sides of the spherulite forced in and the cavity partly filled by the matrix. But this was not liquid enough to enter very far into the hollow cavity, nor has it filled up the cracks on the outside of the shells. It is evident that there was motion in the lava after the large hollow spherulites had formed, and that they were rigid crystalline bodies. It is quite as evident that the delicate lithophysae of various sizes were not formed before the lava came to rest, because they have not been crushed in any case, although their shells are often much thinner than those of the hollow spherulites.[10]

There seems to be no need to postulate shearing to explain the distribution of the spherulites which grew before cessation of flowage (Fig. 2) and which occur in layers parallel to the flowage. The parallelism is reasonably explained by spreading during flow of the lava. The spherulites which grew after cessation of flowage, and which are aligned parallel to the flowage, seem more reasonably explained by growth along certain flow layers rather than along shear planes paralleling the previously formed flow planes, especially where the latter are intricately folded. In this connection it is pertinent to call attention to Iddings' further statement regarding the rhyolite at Natural Bridge, namely, that the "eccentric and irregular shapes [of the lithophysae] are more or less in accord with the crooked and distorted banding which marks the planes of flow in the rock." This suggests that the crystallization of the lithophysae, after emplace-

[9] *Ibid.*, pp. 522–24. [10] *Op. cit.*, p. 386.

ment of the lava, was controlled by the flow banding, not by shear planes. It should be pointed out that the rhyolite in the region near the head of Yellowstone Canyon is traversed by innumerable small shears. These fractures, however, cut across both the flowage and the coarser layering and are clearly of later date.

The evidence so far in hand seems to support the thesis that the greater part of the layering of the rhyolite is due to spreading of a heterogeneous magma. Layering as a result of late-stage shearing may have occurred locally.

The writer wishes to suggest an additional possibility—that some of the layering may locally postdate even final consolidation of the lava. In the neighborhood of Lower Falls, dense, reddish rhyolite, similar to that of the resistant layers, is developed along joint cracks cutting across the layering. The crystallization of this reddish rhyolite was undoubtedly influenced by the joint cracks. It is suggested that the joint cracks permitted the penetration of crystallization-stimulating gases either from still-liquid lava below or from a deeper source. Such gases, rising through the consolidated lava along joint cracks or other passages, may have escaped laterally along the more pervious flow layers and stimulated crystallization within them. The vapors which rise toward the surface in Yellowstone may even now be promoting such crystallization.

SUMMARY

The wavy lamination of the Yellowstone rhyolite is clearly a flowage phenomenon. A similar origin is favored for the type of coarse stratification herein described, although the possibility remains that late-stage shearing may locally have had some influence on its development. A further possibility is suggested—that the crystallization of the dense layers may have been brought about in some cases by vapors which came in after consolidation of the lava, rather than by vapors originally included within it.

ACKNOWLEDGMENTS.—The writer is indebted to R. T. Chamberlin, C. N. Fenner, and H. A. Brouwer for critical reading of the manuscript. He has had the pleasure of discussing some of the problems of the rhyolite with Dr. Brouwer in the field.

ON SHEAR[1] CONTROL OF STRUCTURES OCCURRING IN RHYOLITES OF YELLOWSTONE PARK: A REPLY

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Dr. Howard's discussion of my article on the Yellowstone rhyolites being in its final form the result of a correspondence between us on the subject, there remains only one point which seems to have to be explained more explicitly: flowage phenomena *are* abundant. This fact has never been doubted by anybody. As a matter of fact, a number of slides in my collection are quite of the type illustrated in the Figures 1 and 2 of Howard's discussion. The point, however, is that I thought it interesting to call attention to the fact that other phenomena point to other modes of origin.

A detailed discussion in favor of shear control of the development of lamination in obsidian from Lipari and other lavas is given by Philipp[2] in a paper which was published since my article was in print. He concludes that, during the cooling of a sufficiently fluid flowing melt, a turbulent phase will be successively followed by a continuous laminar and a discontinuous laminar phase before final solidification. The discontinuous laminar phase has analogies with glacier movement.

Continued shearing may promote the escape of volatile constituents and influence the crystallization[3] in rhyolites. Much is still to be learned about the influence of shearing stress on crystallization; and because sufficient theoretical and experimental data are lacking

[1] "The use of the term shear or shearing plane is a source of much confusion" (Leith, *Structural Geology*).

[2] H. Philipp, "Bewegung und Textur in magmatischen Schmelzflüssen," *Geol. Rundschau*, Band XXVII (1936), S. 321–65.

[3] The fact, however, that even for those places where the occurrence of steeply dipping lava flows with lamination and stratification is described, stony rhyolites in isolated bodies of different size are mentioned in my original paper, shows clearly that I too take crystallization, without relation to deformation, into consideration.

the matter cannot yet be presented on systematic lines. For the present the investigations are mainly directed toward definite establishment of the characteristics of the rocks. Considered in space and time, we face a very complicated problem: in a cooling lava flow the rate of cooling, the escape of volatile substances, the resistance in front and below, and the direction and rate of movement, they all may be varying. If lavas cease to flow in an early stage of consolidation and shear has not been effective, then shear control of rhyolite structure can of course not be expected.

REVIEWS

The Birth and Development of the Geological Sciences. By FRANK DAWSON ADAMS. Baltimore: Williams & Wilkins Co., 1938. Pp. 506; figs. 79. $5.00.

Professor Adams, believing with Aristotle that "he who sees things grow from the beginning will have the best view of them," has with this book made it possible for everyone to have the best possible view of the beginnings of geology. In preparing the present history the author has been able, in almost all cases, to go back to the original authorities, and he is of the opinion, which is likely to be shared by most reviewers, that few works of real importance have been missed. Professor Adams' compelling interest in his subject is well shown by the fact that for more than a score of years he has been collecting a library on the history of geology, which now numbers over one thousand volumes, many of which are extreme rarities. In addition, he has had access to most of the really great libraries of North America and Europe. These admirable bibliographic backgrounds are brightly reflected in the book itself, for, through it, the reader can lean over the shoulder of the author and catch scintillations from "many a quaint and curious volume of forgotten lore." The result is so interesting that, as reviewers are accustomed to exclaim regarding potential best sellers, "the reader cannot put the book down until he has read it through."

The Birth and Development of the Geological Sciences is divided into fourteen chapters, whose titles give a somewhat inadequate concept of the scope of the work. Chapter i is an "Introduction" which includes also a statement regarding sources. Chapter ii discusses "The Geological Sciences in Classical Times." Early Grecian writers from Thales and Anaximander to Theophrastus, later Grecians from Agatharchides to Strabo, the Alexandrine lapidaries, and the Romans from Lucretius to Solinus play the chief roles in this forty-two-page section.

Chapter iii is concerned with "The Conception of the Universe in the Middle Ages." Chapter iv, the "Generation of Stones," is a fascinating account of lapidifying juices, digestive organs of minerals, male and female minerals, stones which become pregnant and bring forth young, stones growing in animals and plants, trees whose tears harden into onyx,

and a bewildering assemblage of similar once almost universally believed absurdities. Chapters v and vi trace the history of "Medieval Mineralogy" and the "Birth of Modern Mineralogy" from Agricola to Werner and Berzelius. These two sections involve seventy-odd pages crammed with interesting items. Chapter vii discusses "The Birth of Historical Geology" and is chiefly devoted (1) to Werner, his students, and the Neptunian theory and (2) to Hutton and the Plutonists. Chapter viii is the story of "Figured Stones" and the "Birth of Paleontology."

"The Origin of Metals and Their Ores" is discussed in chapter ix, and views on this subject from Aristotle through Agricola to Werner are considered, the sad conclusion apparently being reached that even today our knowledge of the subject is only a little way farther on "toward what it is not human nature ever perfectly to complete." Chapter x, one of the best in the book, deals with "The Origin of Mountains." Some seventy pages are devoted to the discussion of ideas on mountain-building in classical times, to the medieval speculations of Avicenna, Da Vinci, Dante, and Fallopius, and to the more detailed investigation between 1600 and 1830 made by Steno, Arduino, Lehmann, von Buch, Hutton, De Saussure, and Elie de Beaumont. Chapter xi deals with "Earthquakes and the Character of the Interior of the Earth" and describes the evolution of ideas in seismology. Chapter xii on the "Origin of Springs and Rivers" considers the speculations of a long line of imaginative writers. The works of Athanasius Kircher here receive the attention which they amply merit and never before have obtained. Vallisnieri also comes into proper focus in this chapter, and some of his admirable geological sections are reproduced.

Chapter xiii is rather gratuitously entitled "Quaint Stories and Beliefs," for the concepts chronicled here can scarcely be said to be much "curiouser" than those related in earlier chapters. Chapter xiv, the "Conclusion," contains a backward glance over the long and difficult road traveled to modern concepts, as well as an attempt to ascertain what so commonly "turned the seekers aside from the highroad of knowledge into the bypaths of misunderstanding and error."

Professor Adams has produced a volume which is both better written and more thoroughly documented than Geikie's classical *Founders of Geology*. It is more interesting and, at the same time, somewhat less detailed than Zittel's encyclopedic *Geschichte der Geologie und Paleontologie*. Just as the finest steak may be improved by condiments, however, the enjoyment of Professor Adams' volume can be increased by reading it along with Edna Kenton's all too little known *Book of Earths*

(Morrow, 1928) and George H. T. Kimble's *Geography of the Middle Ages* (London: Methuen, 1938). And if one is interested in really strong relishes, he should add *Moons, Myths, and Man* by H. S. Bellamy (Harper, 1938) and *Lost Atlantis* by James Bramwell (Harper, 1938). For these are, in the main, books of "wonders," and Professor Adams' volume, since it is concerned chiefly with the history of geology up to "about 1825," falls properly into the same category.

Some readers may feel that too much space was devoted to the undeniably fascinating early geological aberrations and not enough to the amazingly astute deductions of a number of the early earth scientists. But on the whole the book is reasonably well rounded even as to the amount of space allotted to the various topics. Paleontology and historical geology, to be sure, receive a less complete treatment than possibly would have been allotted them had the work been prepared by a stratigrapher or paleontologist.

There is, surprisingly, only one reference to Linnaeus and no mention of any of the North Italian school of early micropaleontologists such as Gualterius or Plancus, nor of Van Leeuwenhoek, whose work with the microscope made their observations possible. There is also no consideration of the geologically famous Temple of Jupiter Serapis, none of ancient concepts as to the age of the earth, or of the fable of Atlantis; there is only one reference to Fracastoro, only one to Sedgwick, and Murchison is referred to only because of his reminiscences of von Buch, Smith, and Featherstonhaugh. Berzelius, Mallet, Buffon, and others are not dignified by a first name, and Agassiz, von Alberti, Beyrich, De Blainville, Buckland, Bunsen, Conybear, De la Beche, St. Hilaire, D'Halloy, Phillips, and Sowerby are merely some of the many who are not mentioned at all. Most of these omissions are doubtless intentional, since the book apparently is designed to end when geology, as a science, outgrows its adolescence. Since the various subfields "grew up" at different rates, it is perhaps not surprising that there are different terminal dates for the several chapters, most of which, apparently for contrast's sake, are concluded in somewhat disconcertingly telescoped attempts to bring the subjects discussed up to their modern status. The Index, which Professor Adams doubtless did not personally construct, is incomplete and marred with typographical errors, and, finally, it seems to the reviewer that many more illustrations might easily and profitably have been included.

It has become customary for reviews to list errors in the volume receiving notice, but many of the points just cited involve mere differences of opinion; and, in any case, the demerit marks are inconsequential in

comparison to the innumerable entries on the credit side of the ledger. So much is this so that it is no hyperbole to say that geologists the world around owe Professor Adams a profound debt of gratitude for making available to them, in delightful form, the all too little known foibles of their scientific forefathers.

CAREY CRONEIS

Map 266A, Kenora Sheet, Ontario. 2d ed. Ottawa: Dept. Mines and Resources, Bureau of Geology and Topography, 1939. 1 inch = 8 miles (1/506,880). Lat. 48°00′ to 52°00′ and Long. 90°00′ to 95°10′ (approximately).

The first edition of this map, published in 1933, has been out of print for several years. The appearance of the new map with extensive revisions and additions, including a revised legend and marginal text, is particularly welcome. Dr. Tanton has been especially conservative in the compilation of this map and has not attempted to fit the diverse rocks of the Archean of this region into a set system of classification, with one notable exception. The Coutchiching, of which Dr. Tanton is the chief proponent, is shown more extensively than on any map hitherto published. The volcanic schists (Keewatin), various post-Keewatin sediments (Seine, Steeprock, etc.), a minor volume of basic intrusives, and granites complete the Archean. Noteworthy is the absence of the term Laurentian in the legend, marginal text, or map. Dr. Tanton believes all the granite to be intrusive into the rocks with which it is in contact. A very little later pre-Cambrian (Animikie and Keweenawan) is shown in the southeast corner of the area.

Dr. Tanton and the Bureau of Geology and Topography of Canada are to be congratulated on the compilation and publication of this map which, together with *Map 308A* (Lake Nipigon sheet) and *Map 155A* (Lake Huron sheet), all of the same scale, depicts the state of knowledge of the Canadian part of the Shield most thoroughly studied by geologists.

F. J. PETTIJOHN

"Lexicon of Geologic Names of the United States (including Alaska)," by M. GRACE WILMARTH. *U.S. Geological Survey Bulletin 896.* Washington, D.C., 1938. Part I, A–L, pp. 1–1244; Part II, M–Z, pp. 1245–2396. $2.50 for the set.

The amount of time spent by geologists in looking up definitions and usages of geologic names must be very great in the aggregate. Very often

one does not know just where to look for the needed information, and additional time is consumed in locating the information once the proper volume has been found. Commonly, also, one wishes to know for immediate use the stratigraphic position, geographic location, or some other leading fact about a rock unit but cannot then afford the time to make a search for it.

Now, with the appearance of Miss Wilmarth's lexicon of geologic names, much time will be saved to geologists and much information will be immediately available which otherwise they might try to get along without. For each name the lexicon gives the age and area of occurrence of the rock unit, followed in smaller type by one or more references and a few judiciously selected facts covering the lithology, thickness, underlying and overlying formations, type locality, etc. In the course of time many names have been redefined, some of them more than once. These changes are carefully treated in the lexicon, which is about as nearly up to date as such a huge compilation probably ever can be. The gathering of this information was begun over twenty-five years ago as a piece of "knitting work," or a by-product of regular duties. Like Miss Wilmarth's correlation charts for the different states of the Union, the lexicon will be of great value to the geologic profession.

R. T. C.

Gems and Gem Materials. By E. H. KRAUS and C. B. SLAWSON. 3d ed. New York: McGraw-Hill Book Co., 1939. Pp. xiv+287; figs. 344; 4 colored pls. $3.50.

The arrangement of this book[1] has not varied throughout the fourteen years of its life. The chapters on physical properties, cutting, and polishing of gems and manufactured gems, as well as the descriptions, have been augmented (as they were also in the 1931 edition); a new chapter of three pages covers metals used for gem mountings. There are sixty-five new illustrations, though the total is but nineteen in excess of the last edition.

Rotatory polarization is referred to as circular polarization; the rays of Figure 132 do not obey the law of reflection. The book is intermediate in level between the strictly popular works and one like Eppler's, from which are taken the four colored plates that so enhance the appearance of the present edition. The numerous halftones and the general excellent job of book manufacture will help make the work appeal to a wide group of readers.

D. J. F.

[1] Briefly described in *Journal of Geology*, Vol. XXXIV (1926), pp. 184–85.

Practical Seismology and Seismic Prospecting. By L. Don Leet. New York: Appleton–Century Co., 1938. Pp. 430. $6.00.

The difficulties in writing a book without violating good English usage are legion. It is, therefore, always a pleasure to find a book about the earth in which precise language is used and that, although technical and specialized, is easily read and understood.

Mr. Leet's book is intended as an introduction to seismology, and, as such, much detail and many derivations are omitted. Part I, dealing with the geology of earthquakes, is an interesting and straightforward introduction. Part IV covers the history of seismology and of great earthquakes in an adequate manner. Part V is devoted to applications of seismological methods to prospecting. Curve-fitting, the application of the theory of least squares, and the present patent situation in relation to commercial seismology are considered.

Although the professional seismologist and the experienced seismic prospector may complain of lack of mathematical rigor in Parts II and III (the theory of elasticity and instrumental technique), these portions serve the purpose of introduction. Mr. Leet would be the first to admit their inadequacy for the skilled operator, but would reply that the book is intended for beginners. These two parts give the beginner, who perhaps may read no other work on the subject, as satisfactory an appreciation of seismology as any introductory book in any of the other mathematical sciences.

The reviewer believes that the book may be used as the basis of the seismic part of a year's introductory course in geophysics. The author and the publishers are to be commended upon having produced a satisfactory outline of a subject in which relatively few are sufficiently interested to pursue a college course.

H. W. Straley, III

Engineering Terminology. By Victor J. Brown and Delmar G. Runner. Chicago: Gillette Publishing Co., 1938. Pp. 310. $3.50.

The difficulties of specification writing and interpretation are increased by misunderstandings in nomenclature. Although such misunderstandings are sure to arise, reference to *Engineering Terminology* as authority should decrease litigation and facilitate decisions where it has been initiated.

Brown, Runner, and the publishers are to be congratulated upon the success of their attempt to clarify civil engineering nomenclature. Because about one-sixth of the terms defined are from the earth sciences, the book should be on the shelf of every geologist whose work is connected with that of the civil engineer.

H. W. Straley, III